# THE
# JOURNAL OF GEOLOGY
## A BI-MONTHLY



VOLUME 56

JANUARY-NOVEMBER, 1948



THE UNIVERSITY OF CHICAGO PRESS
CHICAGO, ILLINOIS

THE CAMBRIDGE UNIVERSITY PRESS, LONDON

PUBLISHED JANUARY, MARCH, MAY, JULY, SEPTEMBER,
AND NOVEMBER, 1948

COMPOSED AND PRINTED BY THE UNIVERSITY OF CHICAGO
PRESS, CHICAGO, ILLINOIS, U.S.A.

# CONTENTS OF VOLUME 56

## NUMBER 1



## NUMBER 2

## NUMBER 3



## NUMBER 4

---

NUMBER 5

## NUMBER 6

# *To Our Readers*

FOR fifty-five years the *Journal of Geology* has been under the inspiring and devoted editorships of Thomas Chrowder Chamberlin and Rollin T. Chamberlin. Under their guidance it has become an effective instrument for the dissemination of the results of geological investigations, presenting researches of fundamental importance. Now the editorship passes to Dr. F. J. Pettijohn, of the Department of Geology at the University of Chicago. He will have the active support of a group of leaders in the various fields of geology both in the United States and abroad.

The *Journal* will publish papers reporting fundamental research in all branches of geology. Contributions from any part of the world will be considered and may be written in English, or in French or German with an English summary. To facilitate publication, authors are urged to observe the requirements detailed on the next page.

Among the articles to appear in early numbers of the *Journal of Geology* are the following:

"Megascopic Study and Field Classification of Sediments." By P. D. KRYNINE

"Composition of Meteoritic Matter. III. Phase Equilibria, Genetic Relationships, and Planet Structure." By HARRISON BROWN and CLAIRE PATTERSON

"A Preliminary Report on Vertebrates from the Permian Vale Formation of Texas." By EVERETT C. OLSON

"Cave-in Lakes in the Nabesna, Chisana, and Tanana River Valleys, Eastern Alaska." By ROBERT E. WALLACE

"An Unusual Volcanic Dike and Grooved Lava at Auckland, New Zealand." By J. A. BARTRUM and E. J. SEARLE

## SUGGESTIONS TO CONTRIBUTORS

TEXT—Manuscripts should be written in English, or in French or German with an adequate summary in English. They should be typewritten, on one side of paper only, double spaced, and with wide margins.

ABSTRACT—Each article should contain an abstract which gives a concise summary of the content, including both major and minor points. Abstracts should not exceed 200 words.

REFERENCES—Under the heading "References Cited" at the end of the paper should be listed all literature cited, arranged alphabetically by authors and chronologically under each author, using the following form:

DOE, J. S. (1884*a*) Geology of the Big Ben Mountains: U.S. Geol. Survey Prof. Paper 55, pp. 7–17, pl. 2.
——— (1884*b*) Guidebook for excursions in the Big Ben Mountains, pp. 32, 33, New York, Macmillan Co.
SMITH, A. V. (1926) Discovery of fish remains in Ordovician rocks of the Black Hills (abstr.): Geol. Soc. America Bull. 52, p. 27.
———, and DUNSANY, A. J. (1929*a*) Early Ordovician faunas of South Dakota: U.S. Geol. Survey Bull. 444, pp. 1–76, 3 pls.
———, ——— (1929*b*) Revision of Ordovician stratigraphy of the Black Hills: Jour. Geology, vol. 37, pp. 680–695.

In the text, references to the literature cited should follow this form: (Smith, 1928, p. 36).

FOOTNOTES—Discussion which has a collateral bearing on the subject being discussed but would be too much of an interruption to be put in the body of the text should be inserted as a footnote. Footnotes should be inserted immediately after references are made to them, with lines above and below setting them off from the text. They should be numbered consecutively.

ABBREVIATIONS—This *Journal* uses the abbreviations approved by the United States Geological Survey and listed in the Survey's *Suggestions to Authors*.

ILLUSTRATIONS—Maps, line drawings, and diagrams should be prepared on white drawing paper or on tracing cloth, with India ink. Care should be taken not to overload maps with irrelevant detail or, at the other extreme, to use excessive amounts of space to convey relatively little information. The smallest symbols or letters used should be of such size that they will be not *less* than 1 mm. high after reduction. Unsatisfactory illustrations will be returned to the author or may, with his permission, be redrafted at the editorial office at his expense.

Photographs are reproduced by the collotype process or as halftones. Authors should use the minimum number consistent with adequate presentation of the subject. If a paper includes a disproportionately large number of plates, the editor may ask the author to pay a portion of the cost of illustrations. Photographs should be sharp and clear, printed on glossy paper, and either of the size in which they are to appear in the *Journal* or larger.

Explanations of all figures should be typed on one or more separate sheets. Each figure should be marked for identification and top indicated if there is any doubt.

The original illustrations are destroyed following their publication, unless the author has requested their return in advance of publication.

ALTERATIONS—The cost of excessive alterations (i.e., changes from the original manuscript) made by the author will be charged to him. No changes should be made in proof except errors of typography or of fact based on new information.

REPRINTS—The *Journal* supplies free 50 reprints (without covers) of each major article. This does not include geological notes, discussions, reviews, or communications to the editor. Additional reprints may be ordered when manuscript is submitted or at any time in advance of publication. Rates will be furnished on request.

STYLE—In matters of style, spelling, abbreviation, etc., the two guides to be followed are the University of Chicago Press *Manual of Style* and the United States Department of the Interior *Suggestions to Authors*. Where these authorities differ, the latter reference is to be used.

VOLUME 56 NUMBER 1

# THE JOURNAL OF GEOLOGY

*January 1948*

## EARLY TERTIARY FANGLOMERATE, BIG HORN MOUNTAINS, WYOMING

ROBERT P. SHARP
California Institute of Technology, Pasadena

ABSTRACT

Great accumulations of coarse bouldery gravel along the east base of the central Big Horn Mountains are composed almost wholly of debris derived from the pre-Cambrian core of the range. The name "Moncrief gravel" is proposed for this deposit. It has previously been described as Pleistocene glacial material, late Tertiary to earliest Quaternary bench gravels, or the coarse phase of an early Tertiary basin fill. The gravel was found to be gradational into fine-grained early Tertiary beds and to be separated from the pre-Tertiary rocks of the mountains by thrust faults. For these reasons the Moncrief gravel is identified as an early Tertiary, probably Eocene, fan deposit, formed as the Big Horn Mountains were progressively uplifted and thrust eastward during the Laramide Revolution. Early Tertiary glaciation may have played a part, but this is largely speculation.

### INTRODUCTION

#### GENERAL STATEMENT

Thick deposits of coarse bouldery gravel with interbedded silt, sand, and arkose layers compose large piedmont ridges and spurs along the east front of the Big Horn Mountains. This gravel consists predominantly of granitic rock fragments derived from the pre-Cambrian core of the range and has its best development in Bald, North, and Moncrief ridges (fig. 3), which extend 2–3 miles east from the mountains and rise 1,500 feet above the general piedmont level.

Max Demorest (1938, p. 18) referred to these deposits as the "Clear Creek gravels," but Clear Creek is preoccupied several times over (Wilmarth, 1938, pp. 457–459), so the name "Moncrief gravel," from excellent exposures on Moncrief Ridge (fig. 3) in Sheridan County, is substituted. The best exposures are in high north-facing cliffs at the top of the ridge and on the slopes below; hence the north face of Moncrief Ridge (secs. 34 and 35, T. 54 N., R. 84 W., Sheridan quadrangle) is designated as the type locality. The present topographic surface determines the top of the unit, and its basal and lateral margins are gradational into finer-grained "Wasatch" beds except where the gravel lies with angular unconformity on the Eocene Kingsbury conglomerate and older formations or is in fault contact with pre-Tertiary rocks of the mountains. More detailed work may subsequently provide a better basis of delineation, but for the present the boundaries of the Moncrief may be determined by the lowermost and outermost gravel bed, consisting primarily of pre-Cambrian rock debris. The age

is early Tertiary, with Eocene a strong probability; and the maximum thickness observed on Moncrief Ridge is 1,400 feet.

Brief investigation of these deposits was made in 1940 and 1946, incidental to study of erosion surfaces along the east flank of the Big Horn Mountains. Field time totals 7 weeks.

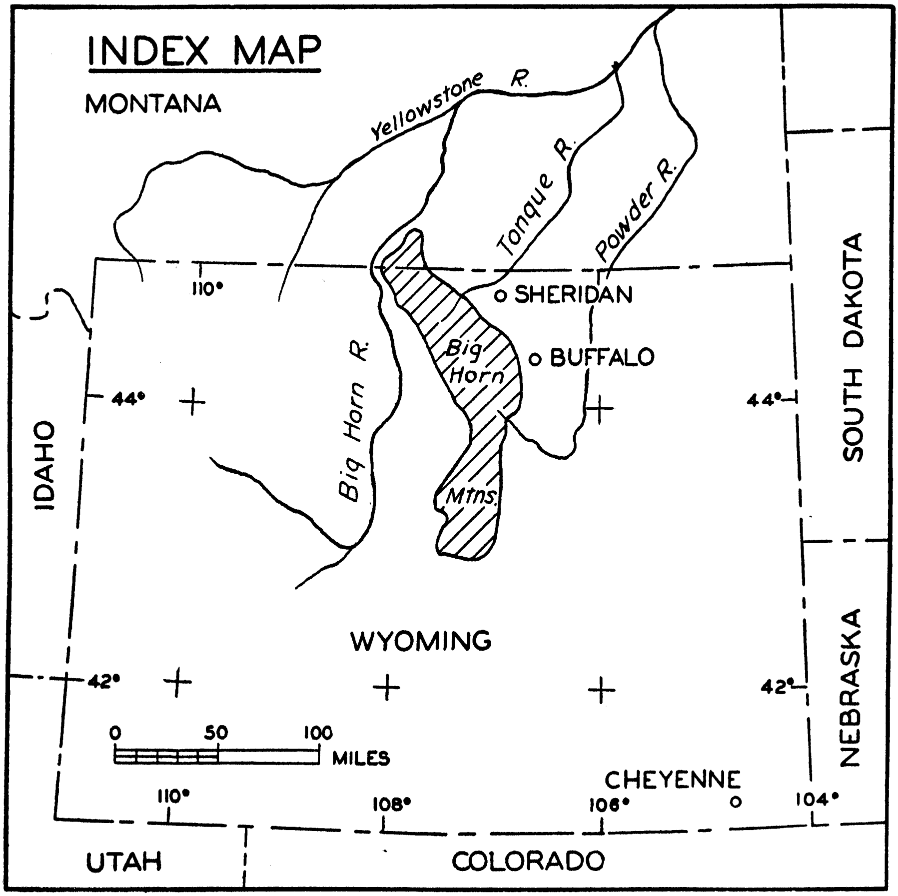



FIG. 1.—Area described lies at east base of Big Horn Mountains between Sheridan and Buffalo

## GEOLOGICAL SETTING

The Big Horn Mountains are the front range of the Middle Rockies in north-central Wyoming (fig. 1). They consist of a major anticlinal uplift trending northwest, with an exposed pre-Cambrian core flanked by upturned Paleozoic and Mesozoic strata. Laramide thrusting is directed eastward in the central section of the range, westward in the north and south sections (Demorest, 1941, pp. 165–166; Chamberlin, 1940, p. 680), and southward at the southern end (Love, 1940, p. 1934; Tourtelot, 1946). In the adjoining basins Mesozoic and Paleozoic rocks are largely buried beneath early Tertiary continental deposits. Topo-

graphically the central part of the range consists of a backbone of high peaks (Cloud Peak, 13,165 feet) flanked by a broad rolling subsummit upland at 7,500–9,000 feet (fig. 2). Imposing parapets developed by erosion of sharply upturned Paleozoic beds form the east front of the range and separate the subsummit upland from the piedmont slope, which lies at 4,500–6,000 feet and truncates tilted Mesozoic and more nearly horizontal Tertiary strata. The backbone of the range has been heavily scoured by late Pleistocene glaciers, which advanced onto the subsummit upland but failed to reach the east base of the range (Darton, 1906*a*, pl. 26).

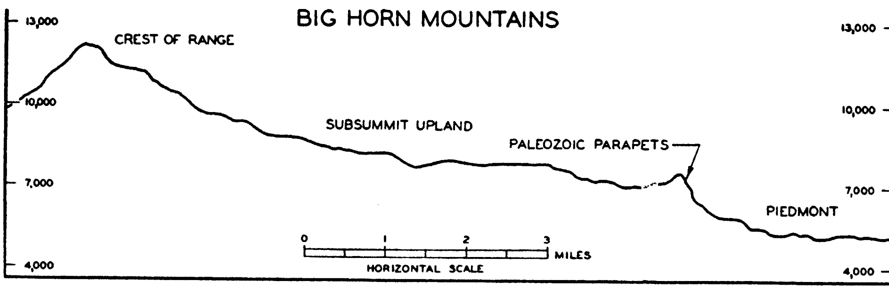



FIG. 2.—Topographic profile, east side of Big Horn Mountains

REVIEW OF PREVIOUS WORK

R. D. Salisbury and Eliot Blackwelder (1903, pp. 221–223) first described the bouldery accumulations at Bald and North ridges and tentatively suggested that they might be glacial deposits of earlier date than the late Pleistocene glaciation recognized higher on the mountains. Blackwelder (1915, pp. 313–315) later compared the deposits of Bald Ridge with coarse fan or outwash gravel on the Black Rock erosion surface (early Pleistocene) in western Wyoming. N. H. Darton (1906*a*, pp. 68–70, pl. 47; 1906*b*, pp. 8–9) mapped all occurrences of the Moncrief gravel and concluded that they were remnants of a late Tertiary or earliest Quaternary bench gravel which at one time extended all along the east-central front of the range. W. C. Alden (1932, pp. 41–44) gives the most thorough treatment of the deposits composing Bald, North, and Moncrief ridges and suggests that they are remnants of great lobate terminal moraines built by early Pleistocene glaciers which flowed from mountains onto the surface of the Flaxville Plain (No. 1 bench). The late Max Demorest (1938, pp. 22–23) suggested two possible interpretations for the Moncrief gravel: (1) It may be a coarse, mountain facies of the widespread Tertiary deposits which filled the intermontane basins and built up the Great Plains, or (2) it is the remnant of local fans deposited in response to local and temporary conditions after the plains surface was largely developed, perhaps in the early Pleistocene. As a result of later field work, he[1] dismissed the second possibility and adopted the view that the Moncrief gravel was gradational into the underlying Eocene Kingsbury conglomerate and also continuous in its uppermost part with Tertiary(?) gravel on the subsummit upland (fig. 2) of the Big Horn Mountains. His published report

[1] Personal letters (1940).

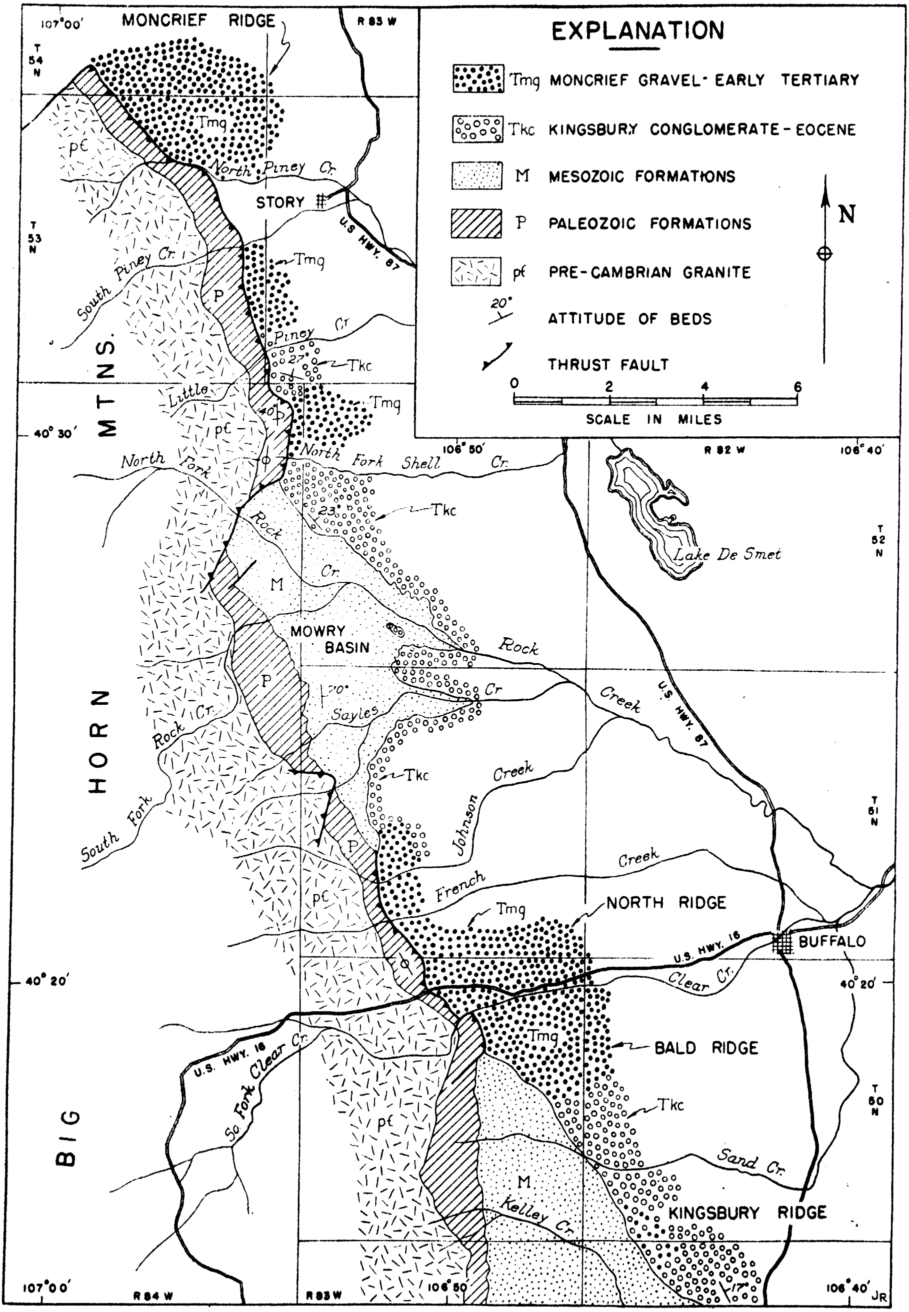



FIG. 3.—Geological map along east base of central Big Horn Mountains, modified from Darton and Demorest.

PLATE I

A

B

*A*, Boulder beds at top of Moncrief gravel in type section on north side of Moncrief Ridge. Arkose, sandstone, and siltstone in middle of exposure. About 200 feet of beds shown.

*B*, Large boulders of pre-Cambrian granite and gneiss in Moncrief gravel north of Little Piney Creek. View northeastward.

PLATE II

A

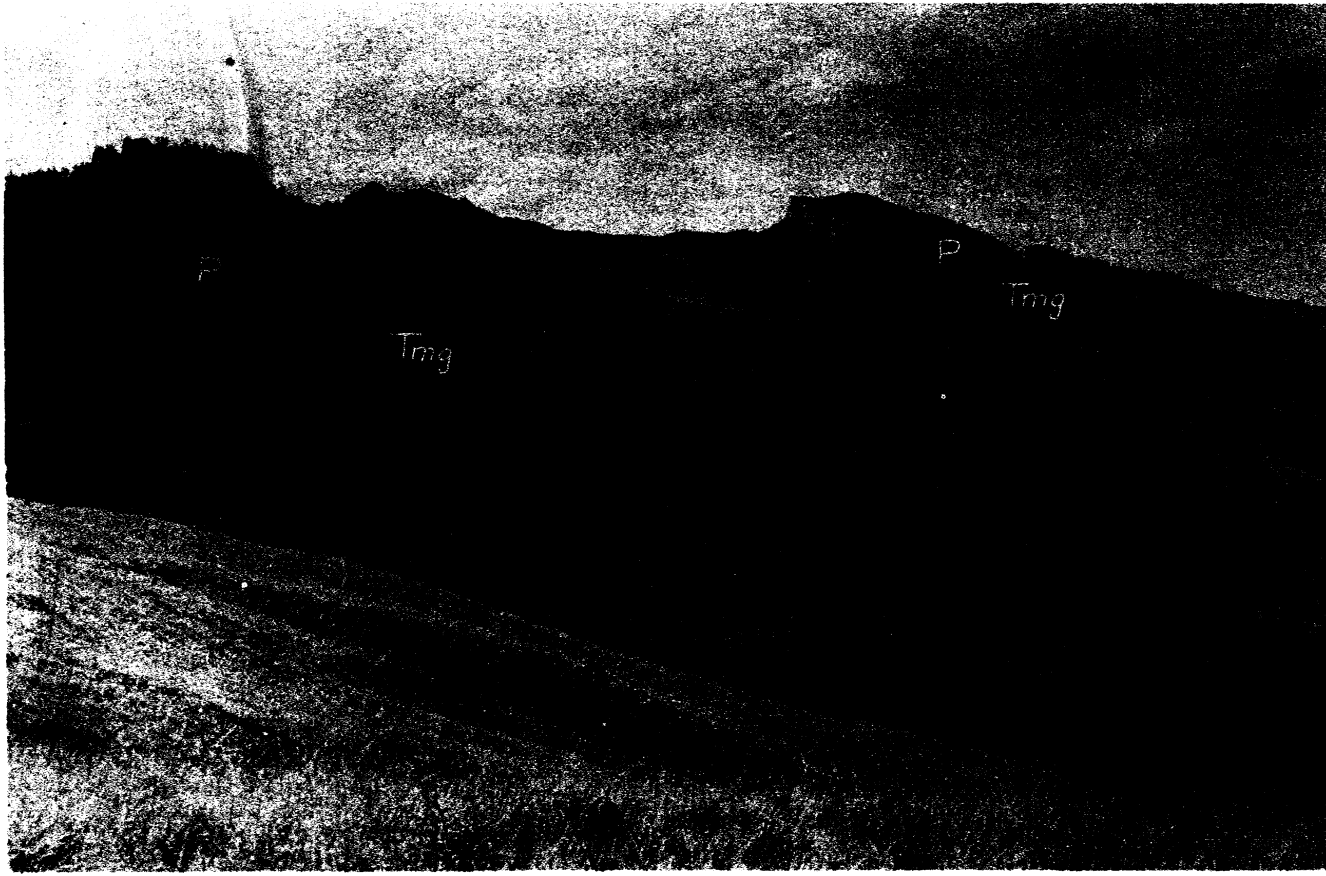
B

*A*, Large boulders of pre-Cambrian gneissic granite from Moncrief gravel on top of Bald Ridge.

*B*, Looking southeast near mouth of Clear Creek with Bald Ridge on left skyline. Moncrief gravel in left half, separated from Paleozoic beds of mountain front by a west-dipping thrust fault.

PLATE III

Perched boulder formed by weathering in fanglomeratic phase of Kingsbury conglomerate, north fork of Rock Creek.




View northward at mouth of French Creek showing Moncrief gravel (*Tmg*) abutting against Paleozoic strata (*P*)

(1941, pp. 167–168), although not explicit, seems to represent the same view. In 1909 J. A. Taff (p. 131) described the gravel composing Moncrief Ridge as gradational into the fine-grained early Tertiary coal-bearing beds of the Sheridan coal field; but his contribution seems to have escaped the notice of subsequent workers. Taff's views and, with some modification, those of Demorest approach most closely the interpretation favored in this study.

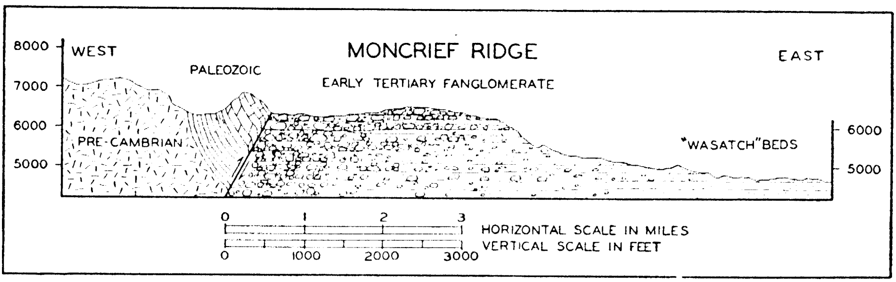



FIG. 4. Sketch showing gradation of early Tertiary fanglomerate (Moncrief gravel) downward and eastward into "Wasatch" beds.

## DESCRIPTION OF THE MONCRIEF GRAVEL

### LITHOLOGY AND CONSTITUTION

This deposit features beds composed of subangular to rounded boulders of pre-Cambrian granitic rocks 1–5 feet in diameter; and larger stones are usually present. For example, on Moncrief Ridge are boulders up to 15 feet long; north of Little Piney Creek (pl. 1, *B*); and between Johnson and French creeks are boulders 25 feet long; south of Little Piney Creek one boulder was measured at 20 × 12 × 10 feet; and on the west end of Bald Ridge is one measuring 27 × 22 × 10 feet. Numerous boulders 10–15 feet in diameter are not exceptional at most exposures (pl. 2, *A*). These larger stones are subrounded, and most are composed of pre-Cambrian granite and gneissic granite. They lie as much as 1 mile east of the mountains and an additional mile from the nearest exposure of pre-Cambrian bedrock. Other lithologies represented, but in smaller amounts, are pre-Cambrian pegmatite, diabase, hornfels, gneiss, and schist, all of which are exposed in the core of the Big Horn Range. Stones derived from Paleozoic formations are extremely rare in the higher beds, but boulders of Paleozoic limestone compose up to 15 per cent of the lower strata, 900–1,400 feet below the top of Moncrief Ridge. The limestone boulders are subrounded to rounded and 6–18 inches in diameter at the most.

Matrix in the gravel is predominantly arkosic and usually sparse, particularly in the upper part where the boulder beds are coarsest, thickest, and most numerous. However, even here fines are not entirely lacking, for beds of arkose and fine micaceous siltstone outcrop in the midst of great boulder layers within 155 feet of the top of Moncrief Ridge (pl. 1, *A*). The deposits become finer downward and eastward, with an increase in number and thickness of arkose and micaceous arkosic sandstone layers. Boulder beds become fewer and thinner and their constituents smaller in the same directions

(fig. 4); and near the base cobble and pebble beds are common.

The finer materials contain crudely spherical arkosic concretions, 1–2 feet in diameter, formed by local cementation. These are identical with concretions in the so-called "Wasatch" of this area. The micaceous sandstone contains smaller limonitic concretions. Cementation of the gravel is poor except locally, where some arkose layers are well cemented and weather out in large fragments or outcrop as ledges. The gravel is white to light-gray and tan. Beds containing stones rich in ferromagnesian minerals are brownish, and micaceous sandy layers are colored grayish-green by chloritized biotite. Bedding is nearly indistinguishable in the coarser phases of the deposit (pl. 1, *A*), and at best it is crude and irregular with many scour channels and some cross bedding in the fine layers. Bits of lignitized woody material are included in fine beds low in the section,[2] and one decomposed lignitized log 1½ feet in diameter was found in the river bank behind the powerhouse on Clear Creek (NE. ¼ of SE. ¼ sec. 1, T. 50 N., R. 83 W.). Disintegration is so far advanced in many exposures that 50–90 per cent of the smaller stones can be chopped away with a geological pick, and some boulders up to 5 feet in diameter are friable to the core. Disintegration is most extensive in the upper 75 feet of the deposit.

Many features of the Moncrief gravel almost exactly duplicate S. H. Knight's (1937, p. 84), description of giant conglomerates at Green and Crook's mountains, Wyoming, and closely resemble conglomerates in the Wind River formation at the south end of the Big Horn Mountains (Tourtelot, 1946).

[2] Demorest (1938, p. 21) also reports coalified wood in these deposits.

### THICKNESS

Thickness of the Moncrief gravel has been variously given as: probably exceeding 100 feet (Salisbury and Blackwelder, 1903, p. 22); at least 300 feet and possibly several hundred feet (Darton, 1906*a*, p. 69); 700–800 feet (Alden, 1932, p. 42); and more than 1,000 feet (Taff, 1909, p. 131). In this study, 485 feet of coarse bouldery gravel were measured by hand-leveling in continuous exposures on the north face of Moncrief Ridge (pl. 1, *A*), and similar boulder beds outcrop at least 900 feet below the top of the ridge on its north side. Along the south side of Moncrief Ridge, gravel beds with pre-Cambrian boulders 2½ feet in diameter outcrop along the bank of North Piney Creek, 1,400 feet vertically below the highest exposure of similar material. New cuts on U.S. Highway 16 along Clear Creek west of Buffalo show that the coarse boulder beds composing North Ridge are at least 1,200 feet thick.[3] It is safe to say that, close to the mountains, gravel composed predominantly of pre-Cambrian boulders is at least 1,200 feet thick on Clear Creek and 1,400 feet thick at Moncrief Ridge.

### DISTRIBUTION

The largest accumulations of Moncrief gravel are in Bald and North ridges, 5 miles west of Buffalo, and in Moncrief Ridge, 2½ miles northwest of Story. However, as shown in figure 3, other sizable deposits of the gravel lie along the base of the range near French, Johnson, Shell, and the several branches of Piney creeks.

Boulders, up to 5 feet in diameter, of pre-Cambrian gneissic granite on Kingsbury Ridge, 4 miles south-southeast of Bald Ridge and 3 miles from the moun-

[3] Demorest (1938, p. 21) recognized that the gravel composing North Ridge was about this thick but gave no figures.

tains, have been mapped by Darton (1906*a*, pl. 47; 1906*b*, areal geology map) and Demorest (1938, fig. 5, p. 18*a*) as part of the Moncrief gravel and are so shown in figure 3. However, this occurrence is regarded with some skepticism because actual exposures of the gravel could not be found and large boulders of Bighorn dolomite (Ordovician) are also present. This material might be coarse debris washed out from Bald Ridge and the mountains during late Cenozoic degradation of the piedmont area.

In all other places Moncrief gravel lies directly against the upturned Paleozoic beds of the range front, which rise several hundred feet higher (pl. 2, *B*, and pl. 4). The gravel attains its greatest elevation, 6,900 feet, in Bald Ridge; and the lowest exposure is along North Piney Creek at 5,100 feet. It extends at least 2½ miles eastward from the mountain front, discounting the questionable occurrence on Kingsbury Ridge.

The subsummit upland of the Big Horn Mountains (fig. 2) is mantled with "Tertiary" gravels, which Demorest (1938, p. 21; ftn. 1) thought were to be correlated with the uppermost beds of the Moncrief. It is not unlikely that the Moncrief gravel formerly overlapped the lower flanks of the mountains, but the gravels on the subsummit upland are thought to be much younger, on the basis of structural and age relations to be detailed shortly. No deposits of the Moncrief were found on the subsummit upland.

Especially worthy of note are the map relations (fig. 3), which show that the Moncrief gravel occurs only where early Tertiary formations lie close to the base of the range. Where such formations are lacking, there is no gravel, and this holds even locally, as in Mowry Basin (fig. 3).

## STRATIGRAPHIC AND STRUCTURAL RELATIONS OF THE MONCRIEF GRAVEL

### WITH THE "WASATCH"

Essentially horizontal beds of arkose and gray-green micaceous arkosic sandstone along the east front of the central Big Horn Mountains are shown on the latest maps (U.S. Geol. Survey, 1925, 1932) as "Wasatch,"[4] presumably on the basis of revisions (Taff, 1909, pp. 127–131; Wegemann, 1917, p. 60; Thom and Dobbin, 1924, pp. 494–498) of Darton's 1906*a*, p. 66; 1906*b*, p. 8) original treatment, in which they were dated as Tertiary or Cretaceous. Layers of identical material are interbedded with the Moncrief gravel even in its highest and coarsest part, and similar fine materials become more abundant downward and eastward in the section. This relation is particularly clear on Moncrief Ridge, where there can be little doubt that the gravel is but a coarse phase of the so-called "Wasatch"[5] (fig. 4). The boulder beds can be seen to lens out eastward along North Piney Creek, so that exposures of the "Wasatch" at the same level several miles east of the mountains consist principally of coarse arkose and micaceous arkosic sandstone. Fine beds in the Moncrief gravel contain coalified plant debris, a further similarity with the "Wasatch," which is coal-bearing in the Sheridan and Buffalo coal fields.

Previous workers, except Taff (1909, p. 131) and Demorest (1941, p. 168; ftn. 1),

[4] R. L. Nace (1936, p. 121) suggests that the name "Wasatch" be avoided as unnecessary and meaningless. Proper identification and naming of Paleocene and Eocene units east of the Big Horn Mountains remains for future detailed paleontological and stratigraphic studies, and the word "Wasatch" as used here is simply a quotation from earlier maps and publications.

[5] Taff (1909, p. 131) recognized and properly interpreted this relation in 1909.

have inferred or postulated an unconformity between the gravel and the "Wasatch," but extended search failed to reveal any discernible break between these units.

WITH THE KINGSBURY CONGLOMERATE

The Kingsbury is a limestone conglomerate, locally fanglomeratic, defined and named by Darton (1906*a*, p. 60) from Kingsbury Ridge (fig. 3). As he mapped the formation, it is a lens-shaped deposit, 2–5 miles wide, extending 40 miles along the east base of the central Big Horns. Wegemann (1917, p. 59) speaks of it as a "fan deposit," and Darton (1906*b*, p. 8) states that it was produced by uplift of the range. Demorest (1938, p. 20) also emphasizes that the Kingsbury is a fanglomerate, formed by uparching of the central segment of the Big Horn Mountains.

One of the best exposures of the Kingsbury is in prominent bluffs northeast of the North Fork of Rock Creek (sec. 19, T. 52 N., R. 83 W., Fort McKinney quadrangle), where it is chiefly a heterogeneous, poorly bedded, massive fanglomerate of subangular to subrounded fragments of Paleozoic limestone up to 5 feet in diameter (pl. 3). No boulders of pre-Cambrian origin were found here, although a few have been seen in other exposures. Cementation, at least locally, is good. Not less than 400 feet of beds are exposed north of Rock Creek; and Darton (1906*b*, p. 8) gives the maximum thickness of the formation as 2,500 feet (fig. 5).

In other localities farther from the mountains, the Kingsbury is predominantly a conglomerate containing well-rounded to subrounded pebbles and cobbles of Paleozoic limestones and occasional stones of pre-Cambrian origin. The Kingsbury is unconformable on tilted Mesozoic beds (Knowlton, 1909, p. 209; Gale and Wegemann, 1910, p. 144; Wegemann, 1917, p. 60; Demorest, 1938, p. 19), although Darton (1906*b*, p. 8) thought this contact was conformable and dated the formation as later Cretaceous. Wegemann (1917, p. 60) placed the Kingsbury in the "Wasatch," but Knowlton (1909, p. 210) later summarized fossil evidence indicating that it is probably Fort Union, and it is so shown on the latest maps (U.S. Geol. Survey, 1925, 1932). Nace (1936, p. 100) includes the Kingsbury in the "Wasatch Group" for purposes of discussion and summarizes the conflicting evidence as to its age, which he tentatively puts as Paleocene(?); but Jepsen (1940, p. 242) cites fossil evidence indicating that the Kingsbury is Early Eocene. Roland W. Brown, of the United States Geological Survey, has collected vertebrate remains from the Kingsbury conglomerate which definitely fix its age as Eocene.[6]

Bedding in the Moncrief gravel, where discernible, is essentially horizontal, with eastward dips not exceeding 2°–3°. All dips measured in the Kingsbury conglomerate are 17°–23° northeastward. The actual contact between the Moncrief and the Kingsbury has not been seen; but, because of this discordance in dips, it is thought to be an angular unconformity, and the marked difference in composition of the two formations also suggests a break between them (fig. 5). This was Darton's (1906*b*, p. 8) original interpretation, as well as Alden's (1932, p. 42); but Demorest (ftn. 1; 1941, pp. 167–168) described this contact on the north side of Moncrief Ridge as gradational, although in earlier work (1938, p. 21) he treated the Moncrief-Kingsbury contact in the area of Clear Creek

[6] Personal letter (1947).

as an angular unconformity. It now appears that the gradation described by Demorest is the Moncrief-"Wasatch" relation treated in the preceding pages.

In places, outcrops of the Kingsbury conglomerate have greater elevations than the lowest near-by exposures of Moncrief gravel. This relation is well shown on Little Piney Creek and is thought to indicate that the Moncrief was deposited on a land surface of considerable local relief (fig. 5). Ridges and hills of this early landscape, held up by firmly cemented Kingsbury conglomerate, have been exhumed in the modern cycle of erosion.

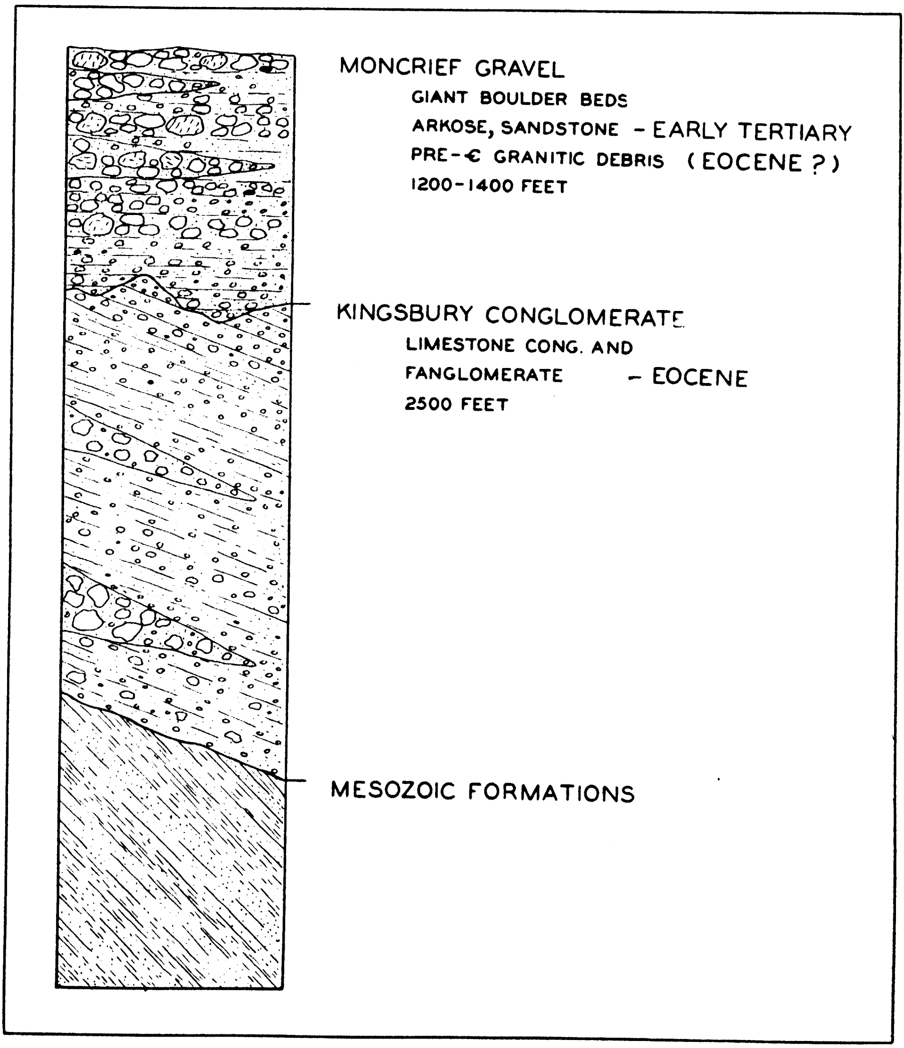



FIG. 5.—Columnar section of Moncrief gravel and Kingsbury conglomerate

WITH PRE-TERTIARY ROCKS

The Moncrief gravel rests on tilted Mesozoic strata with a marked angular unconformity, as recognized by Darton (1906*a*, pl. 47; 1906*b*, areal geology map) and as clearly demonstrated by the field and map relations (fig. 3). The contact between gravel and upturned Paleozoic beds forming the mountain front has also been interpreted as depositional (Salisbury and Blackwelder, 1903; Darton, 1906*a*, pl. 47; Alden, 1932; Demorest, 1941, pls. 1 and 4); but such an interpretation meets insurmountable obstacles. One has but to see the great coarse-boulder layers, composed wholly of pre-Cambrian rock fragments, abutting directly against the upturned Paleozoic limestone beds (pl. 2, *B*, and pl. 4) to realize that the Paleozoic rocks could not possibly have had their present position when the gravel was being deposited, unless the gravel came from the east. This possibility is most remote, for the nearest known source of pre-Cambrian debris to the east is the Black Hills, 145

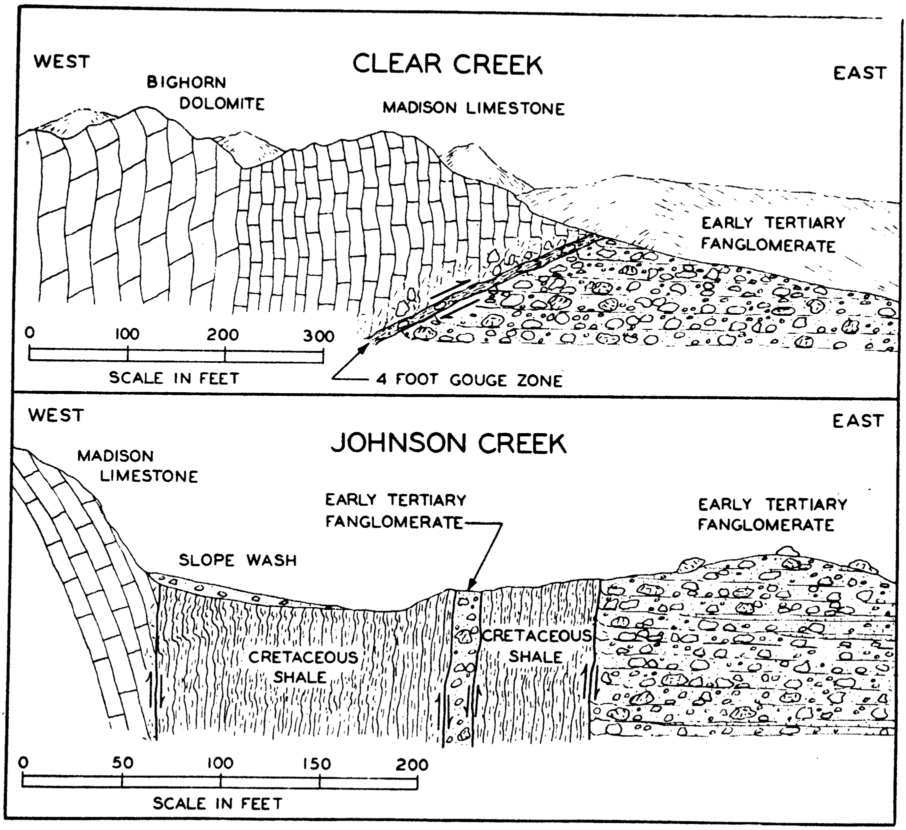



FIG. 6.—Field sketches of fault relations between early Tertiary fanglomerate (Moncrief gravel) and pre-Tertiary rocks of the Big Horn Mountains. Exposure at Clear Creek is in a road cut on U.S. Highway 16. The Johnson Creek exposure is in a gully on the first spur to the north.

miles away. The relations are those expected of a fault; and actual exposures of a fault contact between Moncrief gravel and Paleozoic rocks have been observed at Clear Creek, north of Johnson Creek, and on North Piney Creek. Details of the fault exposures on Clear and Johnson creeks are sketched in figure 6. In the summer of 1946 a new road cut on U.S. Highway 16 along Clear Creek gave a superb exposure of the fault, showing Paleozoic limestone thrust eastward over Moncrief gravel along a plane dipping 25° west. On Johnson Creek the fault is a complex zone, consisting of several nearly vertical planes of displacement involving Paleozoic limestone, distorted Cretaceous shale, and the Moncrief gravel. Before faulting, the gravel probably overlay the Cretaceous shale with an angular unconformity; and in the subsequent displacement a thin sliver or wedge of the gravel was dropped down into the shale, which, in turn, forms a larger slice between the Paleozoics and the gravel. The fault planes appear nearly vertical in the exposure, but they may dip westward at depth. On North Piney Creek the fault is a thrust dipping westward.

On the basis of these exposures and from the fact that Paleozoic strata are steeply tilted or even overturned wherever the Moncrief gravel abuts against them, the contact is interpreted as a fault in all places. Nowhere was a depositional contact observed between these two units. Stratigraphic throw on the fault must have been at least 2,000 feet to give the present relations. Thrust faults have long been recognized (Darton, 1906*b*; Demorest, 1941; Bucher, Thom, and Chamberlin, 1934, p. 169) along the east base of the central Big Horns, but it has not been shown previously that the Moncrief gravel is involved in some of the thrusting.[7] Faulting occurred at more than one period, for Demorest (1941, pp. 165–166) and Darton (1906*a*, p. 93) recognize at least two episodes of faulting. Darton (1906*b*, pp. 60–61, 93) also speaks of the Kingsbury conglomerate–Paleozoic contact as being a fault in some places and an overlap elsewhere. Some skepticism is entertained with respect to a few of Darton's fault relations, as he appears to have mapped great landslide masses as fault blocks.

### AGE OF THE GRAVEL

The Moncrief gravel has previously been dated as early Pleistocene (Salisbury and Blackwelder, 1903, pp. 222–223; Alden, 1932, p. 42), late Tertiary or earliest Quaternary (Darton, 1906*a*; 1906*b*), and has been compared with the Bishop conglomerate (Atwood and Atwood, Jr., 1938, p. 966; Rich, 1910, pp. 601–632) of the Uinta Mountain region, which is thought to be Miocene or Oligocene by W. H. Bradley (1936, p. 163, 185) or about Pliocene by W. W. Atwood (1940, p. 313). Demorest (1941, pp. 167–168) includes the Moncrief beds in his Tertiary(?) gravels, which may range from Eocene well up into the Tertiary. Taff (1909, p. 131) thought the gravel at Moncrief Ridge was equivalent to nearly all his upper member in the Sheridan coal field, which includes the Tongue River member of the Fort Union (Paleocene) and the Intermediate and Ulm coal groups, which may be Eocene (Thom and Dobbin, 1924, pp. 494–498;

[7] The 1947 Bighorn Basin field conference guidebook prepared by the University of Wyoming, the Wyoming Geological Association, and the Yellowstone-Bighorn Research Association, which appeared since this paper was prepared for publication, makes note (p. 95) of the fault between Moncrief gravel and Paleozoic rocks exposed on U.S. Highway 16 along Clear Creek.

Baker, 1929, p. 24; Nace, 1936, pp. 91, 95, 104).

An early Tertiary age is favored in this paper because the gravel appears to be part of the "Wasatch" on two counts. First, it contains typical "Wasatch" material, including coalified plant remains, and passes by gradation downward and eastward into "Wasatch" beds. Second, the gravel occurs only where the "Wasatch" lies along the base of the range. Structural relations also suggest an early Tertiary age, for the gravel is older than some of the eastward thrusting of the Big Horn Mountains, which is presumably Laramide. If the gravel were early Pleistocene, we should have evidence of at least 2,000 feet of Pleistocene thrust-fault displacement—a new and unlikely chapter in the evolution of the Middle Rocky Mountains.

Exact dating of the Moncrief gravel awaits detailed paleontologic and stratigraphic study of early Tertiary beds in the Sheridan coal field. As it now stands, Taff's (1909, p. 131) original description, as modified by more recent work (Thom and Dobbin, 1924; Baker, 1929), makes the lower part of the Moncrief gravel Paleocene and the upper part Eocene; but the evidence of angular unconformity between Moncrief gravel and Kingsbury conglomerate, already described, renders a Paleocene age for any part of the Moncrief unlikely. The resemblance to conglomerates described by Tourtelot (1946) in the late lower Eocene Wind River formation at the south end of the Big Horn Range strengthens the probability that the Moncrief gravel is Eocene.

## ORIGIN

Proponents (Salisbury and Blackwelder, 1903; Alden, 1932) of a Pleistocene glacial origin for the Moncrief gravel have based their arguments chiefly on relations at Bald, North, and Moncrief ridges. However, the equally coarse gravel near Johnson, Shell, and Little Piney creeks is part of the same deposit, and it is noteworthy that none of these latter streams shows the slightest evidence of glaciation in their lower parts. Furthermore, neither Johnson Creek nor Little Piney Creek extends far enough into the range to reach areas of known glaciation (Darton, 1906*a*, pl. 26; 1906*b*, areal geology map). A Pleistocene age is further ruled out by stratigraphic and structural relations already detailed. These relations likewise render untenable Darton's (1906*a*, pp. 69–70; 1906*b*, p. 9) interpretation that the boulder deposits are late Tertiary or earliest Quaternary bench gravels. In this study the present outcrops of Moncrief gravel are interpreted as remnants of early Tertiary alluvial fans built up along the east base of the Big Horn Mountains as they were being uplifted during the Laramide Revolution.[8]

The composition of the Moncrief gravel clearly indicates derivation from the pre-Cambrian core of the Big Horn Mountains; and the crude bedding, angularity, and size of boulders, sparseness of matrix, and occasional fine beds are all consistent with fan deposition (Trowbridge, 1911, pp. 706–747). The huge

[8] This concept was outlined by Professor W. T. Thom, Jr., in a conversation in 1940; and Demorest had essentially the same idea. It was viewed with skepticism during the early stages of this work, until the accumulating field evidence rendered other interpretations untenable. Love (1940) and Tourtelot (1946) report boulder beds in lower Eocene deposits along the south flank of the Big Horn Mountains; and coarse fanlike deposits in early Tertiary beds bordering other Middle Rocky Mountain ranges have been described by the following: Eldridge (1894, pp. 25–26); Westgate and Branson (1913, p. 144); Blackwelder (1915, pp. 106–108); Bauer (1934, pp. 678–679); Love (1939, pp. 60, 106–107); Branson and Branson (1941, pp. 142–143); Van Houten (1944, pp. 179–181).

boulders furnish no objection, for equally large boulders are found on alluvial fans at the east base of the Sierra Nevada (Trowbridge, 1911, pp. 716–717, 740–743). Conditions causing fan deposition were presumably brought about by uplift of the mountains, and this uplift was probably by stages.[9] Initially Mesozoic and Paleozoic strata were exposed in the uparched area; and deposits composed primarily of Paleozoic fragments, such as the Kingsbury conglomerate, were formed along the flanks of the range. Darton (1906*b*, p. 8), Wegemann (1917, pp. 59–60), and Demorest (1938, p. 20; 1941, p. 168) all recognize the Kingsbury as a product of mountain uplift. A subsequent uplift which extended into the piedmont area deformed the Kingsbury mildly and initiated erosion. This was followed by deposition of coarse fan materials, the Moncrief gravel, derived from the mountains which were being raised abruptly above the piedmont, presumably by faulting. At first, the fans consisted of mixed Paleozoic and pre-Cambrian rock fragments; but, as stripping of the Paleozoic rocks became more complete, a larger percentage of the debris was derived from the pre-Cambrian core, until, finally, practically all the detritus was of pre-Cambrian origin.

Increase in coarseness of the gravel upward in the section may have been due to the more massive nature of the pre-Cambrian rocks at depth, to an increasing rate of uplift, or possibly to early Tertiary glaciation in the high parts of the range, like that recognized elsewhere in the Rocky Mountains (W. W. Atwood, 1915, pp. 13–26; W. W. and W. R. Atwood, 1926, pp. 612–622; W. W. and W. W. Atwood, Jr., 1938, p. 961; Drysdale, 1915, pp. 65–66; Scott, 1938, pp. 628–636). C. J. Hares (1926*a*, pp. 174–175; 1926*b*, p. 175; 1926*c*, pp. 175–176) has suggested that somewhat similar deposits along the flanks of various Rocky Mountain ranges are the product of a widespread mid-Tertiary glaciation, but this has apparently not been confirmed by subsequent investigators. The Moncrief gravel is certainly not a till; but it may consist of coarse outwash debris, although there is no direct evidence indicating a glacial source. Alden (1932, p. 43) found one boulder with something like chatter marks; and one small stone with striae-like scratches was collected during the present investigation. An early Tertiary glacial source for the gravel is not established by such meager evidence, and it is not essential to the major thesis proposed. In fact, the map relations (fig. 3) showing that the Moncrief gravel is best developed along those parts of the range uplifted by thrust faults suggest that the gravel was derived from rising thrust blocks as proposed by Knight (1937) and Tourtelot (1946) for similar gravels elsewhere in Wyoming. The fault contact between Moncrief gravel and pre-Tertiary rocks indicates that movement on the faults also occurred after deposition of the gravel.

[9] Love (1939, pp. 116–117) recognizes eight pulsations of the Laramide Revolution in western Wyoming.

## CONCLUSIONS

A series of coarse bouldery fans composed primarily of pre-Cambrian debris was built up along the east base of the central Big Horn Mountains in early Tertiary time, probably the Eocene. This was the section of greatest uplift, and the fan debris was presumably coarser and thicker here than elsewhere. Subsequent faulting thrust the Paleozoic beds of the mountain front eastward against the gravel, and erosion during the remainder of the Cenozoic has gradually etched out

the finer materials, leaving the thickest and coarsest parts of the fan deposits as prominent ridges in the present landscape. At least three episodes of Laramide deformation are indicated: (1) An uplift which produced the Kingsbury conglomerate. Demorest (1941, p. 168) cites evidence indicating that faulting probably occurred during this uplift. (2) A second uplift, also probably attended by faulting, which deformed the Kingsbury and produced the coarse Moncrief gravel. This episode may have been accompanied or closely followed by alpine glaciation. (3) A third, post-Moncrief, period of thrust faulting toward the east in the central segment of the range.

ACKNOWLEDGMENTS.—Numerous residents of the Big Horn Mountain area, particularly U. J. Post, facilitated the field study. Professor W. T. Thom, Jr., has kindly made suggestions in the course of the work and has read the manuscript. Much profit has been derived from the suggestions of S. H. Knight, D. L. Blackstone, J. D. Love, and H. A. Tourtelot. Preparation of manuscript and illustrations was aided by John Townley and Kenneth Thompson working under a research grant from the Graduate School of the University of Minnesota, and this report was written while the author was a staff member of that institution.

## REFERENCES CITED

ALDEN, W. C. (1932) Physiography and glacial geology of eastern Montana and adjacent areas: U.S. Geol. Survey Prof. Paper 174.

ATWOOD, W. W. (1915) Eocene glacial deposits in southwestern Colorado: U.S. Geol. Survey Prof. Paper 95-B.

——— (1940) The physiographic provinces of North America, New York, Ginn & Co.

——— and ATWOOD, W. R. (1926) Gunnison tillite of Eocene age: Jour. Geology, vol. 34.

——— and ATWOOD, W. W., JR. (1938) Working hypothesis for the physiographic history of the Rocky Mountain region: Geol. Soc. America Bull. 49.

BAKER, A. A. (1929) The northward extension of the Sheridan coal field, Big Horn and Rosebud counties, Montana: U.S. Geol. Survey Bull. 806, pt. 2.

BAUER, C. M. (1934) Wind River Basin: Geol. Soc. America Bull. 45.

BLACKWELDER, ELIOT (1915) Post-Cretaceous history of the mountains of central western Wyoming: Jour. Geology, vol. 23.

BRADLEY, W. H. (1936) Geomorphology of the north flank of the Uinta Mountains: U.S. Geol. Survey Prof. Paper 185-I.

BRANSON, E. B., and BRANSON, C. C. (1941) Geology of Wind River Mountains, Wyoming: Am. Assoc. Petroleum Geologists Bull. 25.

BUCHER, W. H.; THOM, W. T., JR.; and CHAMBERLIN, R. T. (1934) Geologic problems of the Beartooth-Bighorn region: Geol. Soc. America Bull. 45.

CHAMBERLIN, R. T. (1940) Diastrophic behavior around the Bighorn Basin: Jour. Geology, vol. 48.

DARTON, N. H. (1906*a*) Geology of the Bighorn Mountains: U.S. Geol. Survey Prof. Paper 51.

——— (1906*b*) Cloud Peak–Fort McKinney folio: U.S. Geol. Survey Atlas, Folio 142.

DEMOREST, MAX (1938) Structural geology of a part of the east front of the Bighorn Mountains near Buffalo, Wyoming, unpublished Ph.D. thesis, Princeton University Library.

——— (1941) Critical structural features of the Bighorn Mountains, Wyoming: Geol. Soc. America Bull. 52.

DRYSDALE, C. W. (1915) Geology of Franklin Mining Camp, British Columbia: Canadian Geol. Survey Mem. 56.

ELDRIDGE, G. H. (1894) A geological reconnaissance in northwest Wyoming: U.S. Geol. Survey Bull. 119.

GALE, H. S., and WEGEMANN, C. H. (1910) The Buffalo coal field, Wyoming: U.S. Geol. Survey Bull. 381, pt. 2.

HARES, C. J. (1926*a*) Glacial origin of the Bishop conglomerate of Wyoming, Colorado, and Utah (abst.): Geol. Soc. America Bull. 37.

——— (1926*b*) What is the Denver formation? (abst.): *ibid.*

——— (1926*c*) Post-Eocene–pre-Miocene glaciation in the Rocky Mountains (abst.): *ibid.*

JEPSEN, G. L. (1940) Paleocene faunas of the Polecat Bench formation, Park County, Wyoming: Am. Philos. Soc. Proc., vol. 83, pt. 1.

KNIGHT, S. H. (1937) Origin of the giant conglomerates of Green Mountain and Crook's Mountain, central Wyoming: Proc. Geol. Soc. America for 1936.

KNOWLTON, F. H. (1909) The stratigraphic relations and paleontology of the "Hell Creek Beds," "Ceratops Beds," and equivalents, and their reference to the Fort Union formation: Proc. Washington Acad. Sci., vol. 11.

LOVE, J. D. (1939) Geology along the southern margin of the Absaroka Range, Wyoming: Geol. Soc. America Special Paper 20.

——— (1940) Thrust faulting at the southern end of the Bighorn Mountains, Wyoming: Geol. Soc. America Bull. 51, pt. 2.

NACE, R. L. (1936) Summary of the late Cretaceous and early Tertiary stratigraphy of Wyoming: Wyoming Geol. Survey Bull. 26.

RICH, J. L. (1910) The physiography of the Bishop conglomerate, southwestern Wyoming: Jour. Geology, vol. 18.

SALISBURY, R. D., and BLACKWELDER, ELIOT (1903) Glaciation in the Bighorn Mountains: Jour. Geology, vol. 11.

SCOTT, H. W. (1938) Eocene glaciation in southwestern Montana: Jour. Geology, vol. 46.

TAFF, J. A. (1909) The Sheridan coal field, Wyoming: U.S. Geol. Survey Bull. 341, pt. 2.

THOM, W. T., JR., and DOBBIN, C. E. (1924) Stratigraphy of Cretaceous-Eocene transition beds in eastern Montana and the Dakotas: Geol. Soc. America Bull. 35.

TOURTELOT, H. A. (1946) Tertiary stratigraphy and its bearing on oil and gas possibilities in the northeastern part of the Wind River Basin, Wyoming: U.S. Geol. Survey Oil and Gas Inv. Preliminary Chart 22.

TROWBRIDGE, A. C. (1911) The terrestrial deposits of Owens Valley, California: Jour. Geology, vol. 19.

U.S. GEOL. SURVEY (1925) Geologic map of Wyoming.

——— (1932) Geologic map of the United States.

VAN HOUTEN, F. B. (1944) Stratigraphy of the Willwood and Tatman formations in northwestern Wyoming: Geol. Soc. America Bull. 55.

WEGEMANN, C. H. (1917) Wasatch fossils in so-called Fort Union beds of the Powder River Basin, Wyoming, and their bearing on the stratigraphy of the region: U.S. Geol. Survey Prof. Paper 108-D.

WESTGATE, L. G., and BRANSON, E. B. (1913) The later Cenozoic history of the Wind River Mountains, Wyoming: Jour. Geology, vol. 21.

WILMARTH, M. G. (1938) Lexicon of geologic names of the United States: U.S. Geol. Survey Bull. 896, pt. 1.

# PHYSIOGRAPHIC DIVISIONS OF ILLINOIS

M. M. LEIGHTON, GEORGE E. EKBLAW, AND LELAND HORBERG

ABSTRACT

The classification proposes minor modifications in Fenneman's divisions and recognizes subdivisions of the Till Plains and Great Lakes sections, based largely on glacial features. The boundaries and characteristic features of the subdivisions are described, and their origin and relations to glacial features and bedrock topography are discussed.

## INTRODUCTION

During the last sixty years increasing attention has been given to physiographic classification; and the broad outlines first conceived by W. M. Davis, J. W. Powell, and others have crystallized into a widely accepted classification for the United States. The establishment of this classification was due largely to comprehensive studies by the late N. M. Fenneman (1914, pp. 84–134; 1928, pp. 261–353; 1931; 1938). In these studies it was recognized that, with the progress of topographic and geologic mapping and the advances in geomorphic knowledge, numerous refinements and revisions would be made. It is the purpose of the present report to make such adjustments of Fenneman's regional boundaries in Illinois as are warranted by present information and to delineate smaller subdivisions which distinguish physiographic differences that can be shown on large-scale maps and which can be used in regional studies within the state (table 1 and fig. 1). These subdivisions, in turn, may be broken down later into still smaller units which may be used as a basis for description in quadrangle reports and other local studies. Local physiographic areas of this type have been described by F. M. Fryxell (1927, pp. 1–53) and by H. B. Willman (1942, pp. 31–37). In the present paper only the distinctive characteristics of the larger subdivisions are summarized.

The report is the outgrowth of geomorphic and glacial investigations carried on by Leighton since 1919, by Ekblaw since 1923, and of recent studies of the bedrock topography and subsurface Pleistocene deposits by Horberg. Copies of a preliminary draft of the map of physiographic divisions of Illinois were prepared in 1944 and furnished to the Committee on Drainage Basins and Flood Control of the Illinois Post-war Planning Commission and to the Illinois Legislative Flood Investigating Commission.

## GENERAL DESCRIPTION AND REGIONAL RELATIONS

Illinois is essentially a prairie plain, and, compared with many other states, it presents few striking physiographic contrasts. The relief over most of the state is moderate to slight and is not sufficient to exert a marked effect on climate. Situated in the south-central part of the great Central Lowland (fig. 2) and near the confluence of major lines of drainage, it is the lowest of the north-central states. The mean elevation is about 600 feet above sea-level, compared with 700 feet for Indiana, 1,050 feet for Wisconsin, 1,100 feet for Iowa, and 800 feet for Missouri (Gannett, 1892, p. 289). The total relief of the state is 973 feet, the highest point, 1,241 feet above sea-level, being Charles Mound in the northwest corner of the state, and the

lowest point, 268 feet above sea-level, the junction of the Ohio and Mississippi rivers. The greatest local relief is near the major valleys, especially within the driftless uplands of northwestern and southern Illinois, and reaches a maximum of 775 feet between Williams Hill, 1,065 feet above sea-level, and the Ohio River Valley, 290 feet above sea-level, in Pope County of southern Illinois. The local relief in most counties in the state, however, is less than 200 feet.

Although large-scale relief features are absent, the physiographic divisions of the state are readily apparent and assume great local significance. More than nine-tenths of the state lies within the Central Lowland, all of which is glaciated except the Wisconsin Driftless section in northwestern Illinois. The other provinces—Ozark Plateaus, Interior Low Plateaus, and Coastal Plain—lie almost entirely outside the glacial boundary in southern and southwestern Illinois.

TABLE 1

PHYSIOGRAPHIC CLASSIFICATIONS OF ILLINOIS

| Classification by Fenneman | Classification by Leighton, Ekblaw and Horberg |
|---|---|
| Central Lowland province | Central Lowland province |
| Great Lake section | Great Lake section<br>Chicago Lake Plain<br>Wheaton Morainal Country |
| Till Plains section | Till Plains section<br>Kankakee Plain<br>Bloomington Ridged Plain<br>Rock River Hill Country<br>Green River Lowland<br>Galesburg Plain<br>Springfield Plain<br>Mount Vernon Hill Country<br>Dissected Till Plains section |
| Wisconsin Driftless section | Wisconsin Driftless section |
| Ozark Plateaus province | Ozark Plateaus province<br>Lincoln Hills section |
| Salem Plateau section | Salem Plateau section |
| Interior Low Plateaus province<br>Shawnee section | Interior Low Plateaus province<br>Shawnee Hills section |
| Coastal Plain province | Coastal Plain province |

#### FACTORS DETERMINING PHYSIOGRAPHIC CONTRASTS

The physiographic contrasts between various parts of Illinois are due to the following factors and conditions: topography of the bedrock surface; extent of the several glaciations; differences in glacial morphology; differences in age of the uppermost drift; height of the glacial plain above main lines of drainage; glaciofluvial aggradation of basin areas; and glaciolacustrine action.

#### TOPOGRAPHY OF THE BEDROCK SURFACE

Recently acquired knowledge of the topography of the bedrock surface of Illinois, so widely obscured by the glacial

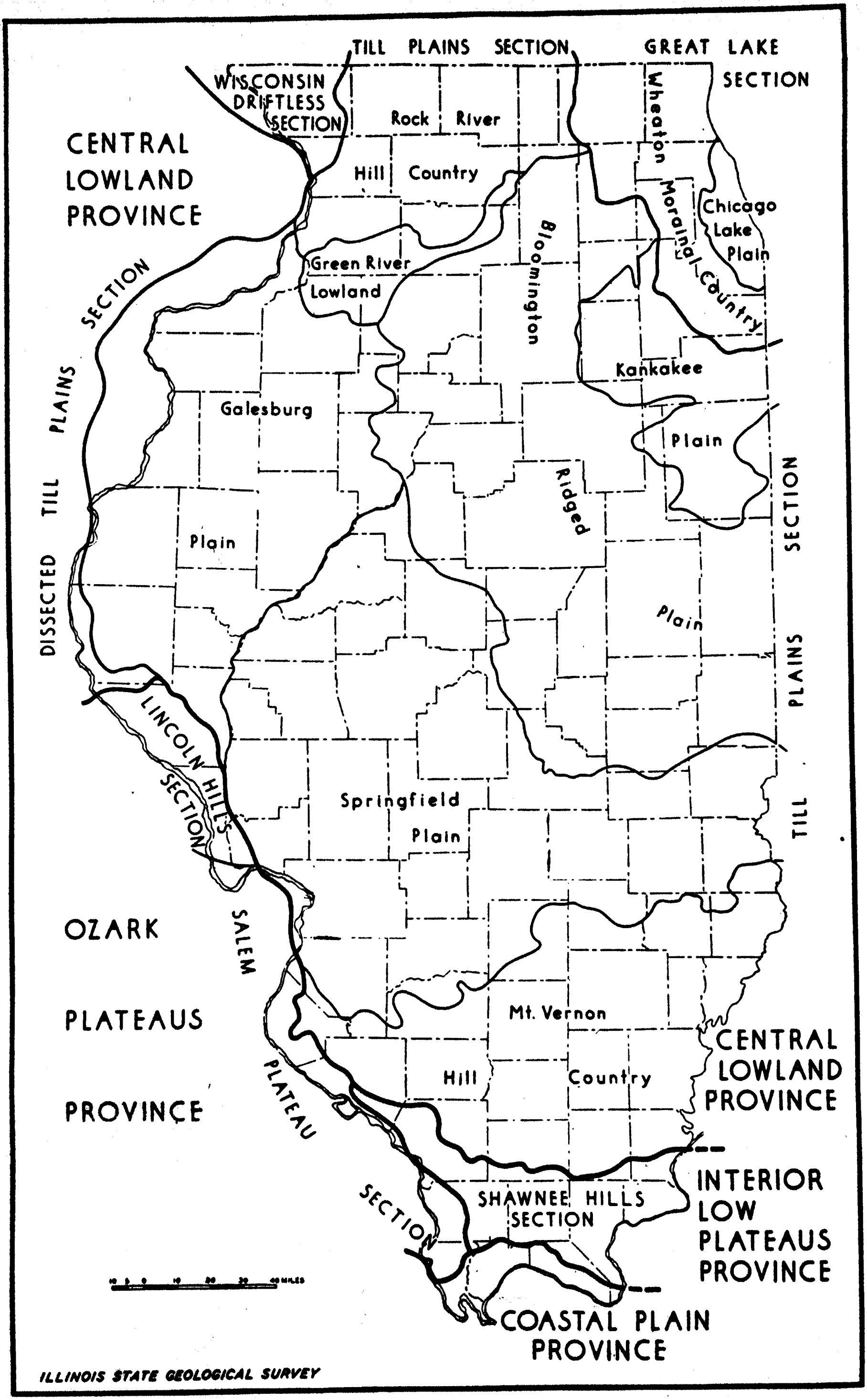



FIG. 1.—Physiographic divisions of Illinois, by Leighton, Ekblaw, and Horberg

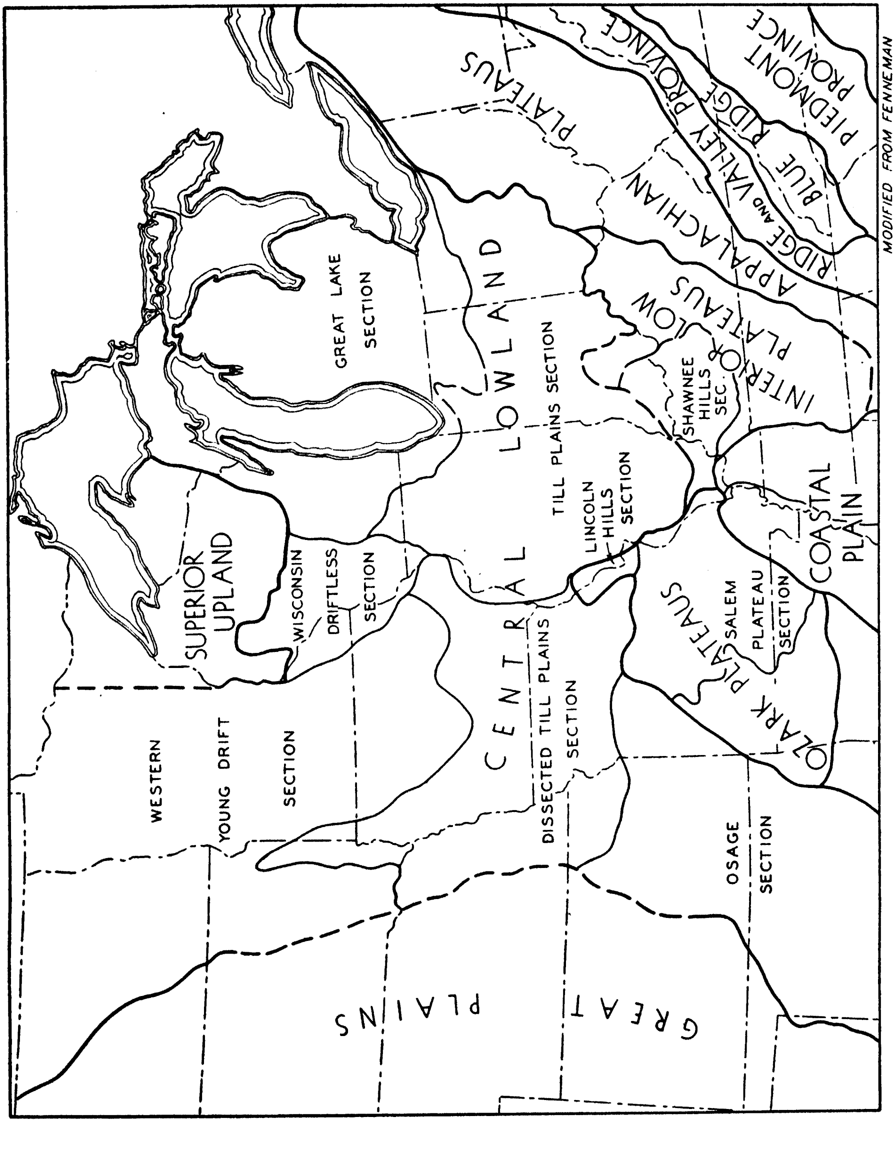



FIG. 2.—Major physiographic divisions in central United States (after Fenneman)

deposits, emphasizes the great importance of this factor in the shaping of the broad physiographic features of the state (Horberg, 1946, pp. 179–192). Prior to glaciation an extensive lowland, eroded on the weak Pennsylvanian rocks of the Illinois basin, covered most of central Illinois. Bordering it on the north, west, and south were uplands that had been developed, for the most part, on the more resistant older Paleozoic limestones and dolomites. This ancient pattern is reflected to an important degree in the present land surface. The extensive lowland of central Illinois provided conditions for the thickest accumulation of glacial deposits and the development of the prairie plains of this portion of the state. The higher uplands of northwestern Illinois and of southern Illinois (the Shawnee Hills) prevented the further movement of the potent Illinoian glacial lobe and caused striking physiographic juxtapositions.

### EXTENT OF THE SEVERAL GLACIATIONS

The effect of the extent of several glaciations is most apparent in the physiographic contrasts between the glaciated and nonglaciated areas of the state—the Wisconsin Driftless section and the bordering Rock River Hill Country in the northwest and the Shawnee Hills section and the Mount Vernon Hill Country in the south. This emphasis upon glaciation, however, should be modified by that measure of difference which already existed in the preglacial landscape. Even so, there is no denying the fact that a topographic revolution resulted from glaciation.

The successive superposition of younger drift-sheets upon older in some parts of Illinois and the deposition of only one drift-sheet in other parts also produced physiographic contrasts in the present topography (fig. 3). The bedrock control that characterizes the Mount Vernon Hill Country and most of the Rock River Hill Country is due to the fact that these regions have but one glacial mantle (the Illinoian), whereas the remainder of the glaciated area of the state was buried beneath either two or three drift-sheets. Nebraskan, Kansan, and Illinoian drifts are present in western Illinois, the first two from the Keewatin field; Kansan, Illinoian, Wisconsin, and possibly Nebraskan drifts from the Labradorean field are present in northeastern Illinois; and Kansan and Illinoian drifts (Labradorean) are present in west-central and south-central Illinois.

### DIFFERENCES IN GLACIAL MORPHOLOGY

Two contrasting types of topography—the Wheaton Morainal Country and the Bloomington Ridged Plain—impressively illustrate differences in glacial morphology. When the Wheaton Morainal Country was formed, the ice lobe was more confined to the deep Lake Michigan basin, and the moraines are closely huddled together. In molding the Bloomington Ridged Plain, the glacier was more extensive and more widely radiating, and during its receding and readvancing substages it formed moraines widely spaced, alternating with nearly featureless ground-moraine plains. The moraines are also generally smoother than the bold moraines of the Wheaton Morainal Country.

### DIFFERENCES IN AGE OF THE UPPERMOST DRIFT

Physiographic contrasts have also been produced by differences in age of the uppermost drift, as in the case of the Bloomington Ridged Plain compared with the Springfield Plain or the Gales-

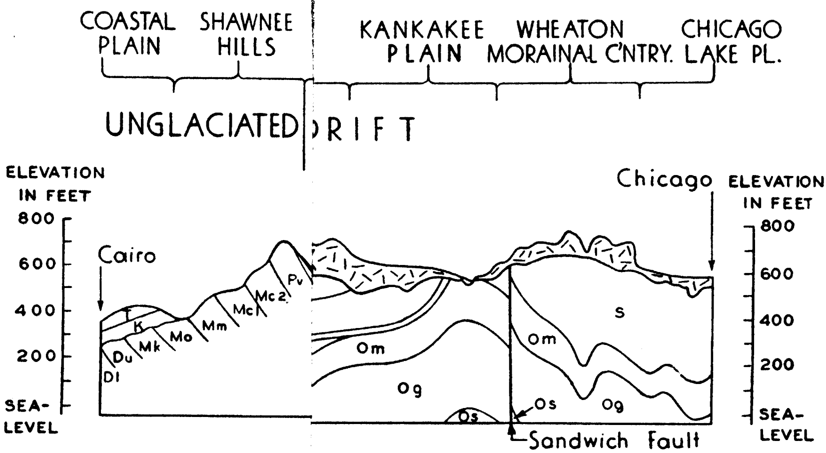



FIG. 3.—Diagrammatic cross sect and bedrock surfaces, and (4) succession and structure of
bedrock formations. Note that the r

*T* —Tertiary syst
*K* —Cretaceous sy
*Pm2*—McLeansboro

*Om* —Maquoketa formation
*Og* —Galena and Platteville formations
*Os* —St. Peter sandstone

burg Plain. The more youthful age of the Bloomington Ridged Plain is responsible for the preservation of its glacial landscape; the greater age of the other two is responsible for their erosional features.

#### HEIGHT ABOVE MAIN LINES OF DRAINAGE

The effect of height above main lines of drainage on physiographic contrasts is the basic factor in the differentiation of the Galesburg Plain from the Springfield Plain. The Galesburg Plain is sufficiently higher above the Illinois River so that its valleys are more sharply incised than were those of the Springfield Plain, and they have different stream regimens.

#### GLACIOFLUVIAL AGGRADATION OF A BASIN AREA

The Green River Lowland is an example of the glaciofluvial aggradation of a basin area.

#### GLACIOLACUSTRINE ACTION

The Chicago Lake Plain is an example of glaciolacustrine action.

## PHYSIOGRAPHIC DIVISIONS

### GREAT LAKE SECTION

The Great Lake section of the Central Lowland province is separated from the Till Plains section to the south because of the bold encircling moraines of Lake Michigan basin, the greater prominence of lakes, and the extent of lacustrine plains in this area. In Illinois a purely topographic boundary between the two sections was drawn by Fenneman along "the outer edge of certain late Wisconsin moraines," which, south of the Kankakee River, form "the western rim of the 'Kankakee swamp . . . .' " (1928, p. 315). In this report the boundary is drawn as follows: Along the west border of the Tazewell Marengo ridge (fig. 4) and the Hampshire ridge south from the Wisconsin-Illinois state line to a point in central Kane County where the Elburn moraine turns westward and southwestward, easterly along an arbitrary topographic boundary to the Cary Valparaiso moraine in western Du Page County, along the outer edge of the Valparaiso moraine southward to the Will County line, and along the outer edge of the closely related Rockdale-Manhattan moraine southward and southeastward to the Indiana-Illinois state line. This boundary coincides, in general, with Fenneman's. The only significant difference is in eastern Kankakee County, where the "Kankakee swamp" is excluded from the section, because the area is an alluviated glacial drainageway and lacks the characteristics of the true lacustrine plains found elsewhere in the section (Ekblaw and Athy, 1925, pp. 417–428). A similar revision of the boundary in the adjoining part of Indiana is implied.

The section is subdivided into the Wheaton Morainal Country and the Chicago Lake Plain, the boundary being marked by the highest (Glenwood) shoreline of glacial Lake Chicago.

*Chicago Lake Plain.*—The Chicago Lake Plain is the well-known featureless "prairie" of early writers of the area. It is characterized by a flat surface, underlain largely by till, which slopes gently lakeward and is interrupted by low beach ridges, morainic headlands and islands, and by two large glacial drainageways along the Des Plaines River and Sag Channel (Bretz, 1939, pp. 43–59). The beach ridges include well-developed spits and bars, which parallel the lake shore or funnel southwestward toward the outlet channel along the Des Plaines River. Blue Island, a prominent morainic island, is situated just east of the outlet channel and rises about 50 feet above

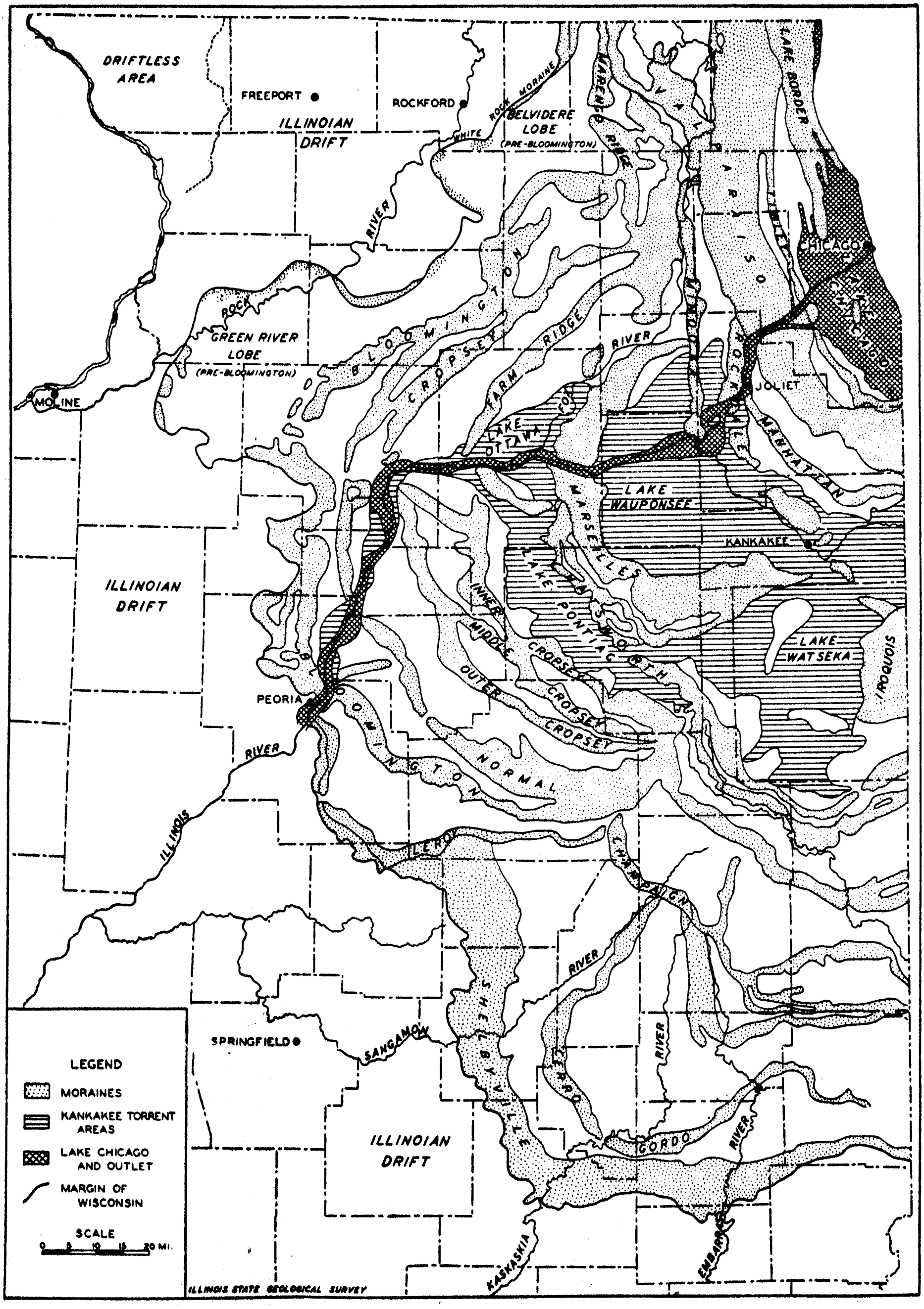



FIG. 4.—Glacial map of northeastern Illinois

the plain. Dunes, which are so conspicuous along the lake shore farther east, are scarcely recognizable and are found in only a few scattered localities.

Originally much of the lake plain was swampy and poorly drained. Its three rivers—the Chicago, the Calumet, and the Des Plaines—are without true valleys and have courses determined largely by beach ridges.

*Wheaton Morainal Country.*—The Wheaton Morainal Country is characterized by glacial morainic topography (mostly of the Cary substage), which is more complex in detail and has more lakes and swamps than do the open stretches of the adjoining Bloomington Ridged Plain. It includes a series of broad parallel morainic ridges, which encircle Lake Michigan. In detail the topography is complicated by a variety of elongated hills, mounds, basins, sags, and valleys. The area is dominated by the Valparaiso moraine, which has the highest elevation and, except where interrupted by valleys, is continuous from Wisconsin to Indiana. With the exception of the Tinley moraine, all other moraines are discontinuous geographic features—those in front of the Valparaiso moraine are overridden by it and those behind are either interrupted by the Chicago Lake Plain or merge with ground moraines. Kames, kame terraces, kettles, basins, and eskers, although not abundant, occur more commonly than elsewhere in the state. Fox Lake and associated lakes are conspicuous water bodies. Small basins of extinct lakes and ponds underlain by stratified silts and clays are found throughout the area.

The topography is determined essentially by thick Wisconsin drift of the Lake Michigan ice lobe, the deposits of which completely buried the underlying bedrock surface. Older Illinoian drift occurs below the Wisconsin deposits along the outer margin of the section in Kane and McHenry counties; but elsewhere this older drift appears to have been removed before the deposition of the Wisconsin drift (fig. 3).

Postglacial erosion has been slight and is restricted largely to youthful valleys along the Fox and Des Plaines rivers. Locally along the Des Plaines Valley near Lemont the Valparaiso drift-sheet is thin, and its major features are determined by a pre-Valparaiso valley system eroded into older underlying drift (Bretz, 1939, p. 52).

*Geologic history.*—The outer moraines in the Great Lake section, including the Marengo, Hampshire, and Cropsey, were formed during the Tazewell substage of the Wisconsin glaciation, whereas the Valparaiso, Tinley, and Lake Border moraines were deposited during the succeeding Cary substage. The Fox River Valley and the valleys of the east and west forks of Du Page River were trenched, apparently by torrential waters, before the advance of the Cary ice. During glacial recession, large blocks of stagnant ice became buried and, upon melting, left the large basins now occupied by Fox Lake and associated lakes.

As the Cary glacier withdrew into the basin of Lake Michigan, Lake Chicago was impounded behind the Tinley and Lake Border moraines, and its discharge waters eroded the outlet channel along the Des Plaines Valley. With retreat of the ice beyond the Straits of Mackinac, this lower eastward outlet into the Atlantic Ocean drained Lake Chicago, and the Chicago Lake Plain emerged with essentially its present aspect. Unlike the beaches north of Milwaukee, the beaches of the Chicago Lake Plain were not subsequently tilted by differential uplift of the region.

TILL PLAINS SECTION

The Till Plains section, which covers about four-fifths of Illinois, is by far the largest physiographic division in the state, and in it seven subdivisions are recognized: the Kankakee Plain, Bloomington Ridged Plain, Rock River Hill Country, Green River Lowland, Galesburg Plain, Springfield Plain, and the Mount Vernon Hill Country. The section is characterized by broad till plains, which are uneroded or in a youthful stage of erosion in contrast with the maturely eroded Dissected Till Plains on the older drift-sheets to the west. The outer boundary coincides closely with the margin of the Illinoian drift.

*Kankakee Plain.*—The Kankakee Plain is a level to gently undulatory plain, with low morainic islands, glacial terraces, torrent bars, and dunes. It is partially fluviolacustrine in origin, but it differs from the lake plains of the Great Lake section in that the lakes which covered it were temporary expansions of glacial floods and did not extensively alter its surface either by deposition or by erosion, except along the courses of strong currents. It could be considered a modified intermorainic basin, floored largely with ground moraine and bedrock.

The district is enclosed by the Iroquois, Manhattan, and Minooka moraines of Cary age on the east and northeast and by the Marseilles and Chatsworth moraines of Tazewell age on the west and south.

Most of the region is poorly drained by shallow low-gradient streams which follow constructional depressions. The two major streams—the Kankakee and the Des Plaines—occupy glacial sluiceways, which, near Kankakee and Joliet, are entrenched in Silurian dolomites. The drift is thick to thin and in the Kankakee region scarcely conceals the bedrock surface.

*Bloomington Ridged Plain.*—The Bloomington Ridged Plain includes most of the Wisconsin moraines of Tazewell age and is characterized by low, broad morainic ridges with intervening wide stretches of relatively flat or gently undulatory ground moraine. In many places the major moraines rise with gentle slopes, and, although they are conspicuous from a distance, they become less so near at hand, whereas the minor moraines are prominent locally. It was in this district more than in any other that the grass-covered stretches of rolling prairie and extensive swamps, described by early settlers, were most typically and extensively developed (Poggi, 1934, pp. 1–124). The outer boundary of the district follows the outer border of the Shelbyville moraine from the Indiana line westward and northward to the Green River Lowland, beyond which it follows the outer border of the Bloomington moraine to its intersection with Hampshire Ridge of the Great Lake section.

The moraines form a series of loops roughly concentric with the outer boundary of the district and from south to north include the following moraines: Shelbyville, Cerro Gordo, Champaign, Leroy, Bloomington, Normal, Outer Cropsey, Middle Cropsey, Inner Cropsey, Farm Ridge, Chatsworth, Marseilles, and Iroquois. The outer moraines are more widely spaced than the inner moraines, and a series of re-entrants in the south-central part of the district indicate an interlobate relationship of the Lake Michigan and Lake Erie lobes. A temporary lake, Lake Ancona (Willman and Payne, 1942, pp. 213–215), is known to have existed behind the Inner Cropsey moraine, but it did not modify the

ground moraine to any important degree.

The glacial deposits are relatively thick throughout the district and completely conceal the bedrock topography, except locally. Illinoian and older drift are present below the Wisconsin in most places, so that the level aspect of present drift-plains is due largely to the presence of the older drift-sheets, which filled in and covered the irregularities of the bedrock surface (fig. 3).

Drainage development is generally in the initial stage, and most streams follow and are eroding in constructional depressions, many of which cross morainic ridges. Undrained basins are much less numerous than in the Wheaton Morainal Country and occur mainly along the morainic ridges. The valleys of principal streams are larger and more numerous than in the Great Lake section, owing in part to greater areal extent of this division and to somewhat greater age, and they have floodplains bordered by valley-train terraces. The Illinois River, the master-stream of the district, has a broad flat-bottomed valley with steep walls and is bordered by numerous narrow steep-walled valleys with steep gradients. Between the "Big Bend" and Peoria the valley coincides with the large pre-Wisconsin valley of the ancient Mississippi and is wider and has a lower gradient than the upstream part of the valley, which is much younger.

*Green River Lowland.*—The Green River Lowland is a low, poorly drained plain with prominent sand ridges and dunes. It is bounded on the north and south by the Shelbyville moraine of the Green River lobe and on the east by the abrupt front of the Bloomington moraine (Leighton, 1923, pp. 265–281). Most of the district is modified outwash plain related to the Bloomington moraine, and it is only in the western part of the area that it merges with the Cary valley-train of the Rock River. Some of the sand ridges are in part bars on the outwash plain, but many are true longitudinal dunes with a west-northwest orientation or crescentic parabola dunes. North of Geneseo, remnants of the Shelbyville terminal moraine can be recognized. At the close of the Cary substage the lowland was a great swamp in which the two principal rivers, Rock River and Green River, flowed sluggishly along poorly defined valleys choked with outwash.

The present lowland coincides in large part with a broad bedrock lowland which was occupied by the Mississippi River up to the time of Wisconsin glaciation; and a remnant of the old southern valley-wall forms a prominent bluff on the south side of the present lowland.

*Rock River Hill Country.*—The Rock River Hill Country is characterized by subdued rolling hill-lands in the stage of late youth to early maturity. It includes the eroded Illinoian drift-plain north of the Shelbyville moraine and Meredosia Valley and a fringe of early Wisconsin drift which lies west of Marengo Ridge.

The Illinoian drift is thin throughout most of the district and is not known to be underlain by older till. Thus the major uplands and valleys are determined primarily by the bedrock surface. The Illinoian drift is without marked ridging, and constructional forms are very localized. In the western part of the district, where it borders the Mississippi Valley, thick deposits of loess and fine sand occur as broad ridges, paha, and dunes on the Illinoian till plain.

The major streams flow radially from a central upland into the Mississippi River on the west and the Rock River on the east and south. Their valleys are relatively broad and steep walled and have

terrace remnants of alluvial fill. The Mississippi River and the upper part of the Rock River occupy large alluviated valleys. Below the mouth of Kishwaukee River, the Rock has cut a post-Illinoian rock gorge, which extends south to the Green River Lowland. Numerous smaller rock gorges are also present along tributaries which locally are superimposed on spurs of the bedrock upland (Hershey, 1893, pp. 314–325). Most of the minor streams are narrow and V-shaped.

*Galesburg Plain.*—The Galesburg Plain in western Illinois includes the western segment of the Illinoian drift-sheet. The till plain is level to undulatory with a few morainic ridges and is in a late youthful stage of erosion. It is bounded by Meredosia Valley and the Wisconsin drift border on the northeast, by the Illinois Valley on the southeast, and by the Illinoian drift boundary on the southwest. On the northwest a continuation of the district across the Mississippi River into Iowa is implied. Four morainic ridges can be recognized in the district—two occur near the drift border, a third lies near and roughly parallel to the Illinois Valley, and the fourth—the Buffalo Hart moraine—extends northward through the central part of the region. Locally, the Buffalo Hart moraine is a prominent physiographic feature.

The district is drained by streams which flow from a central upland westward into the Mississippi River and eastward and southward into the Illinois River. The larger valleys are steep walled, alluviated, and terraced, except for local narrowing along postglacial gorges. Much of the district is relatively high above baselevel, so that the minor valleys are numerous, deep, and youthful.

The Illinoian drift is generally thick and is underlain by extensive Kansan and Nebraskan deposits, especially along buried preglacial valleys. Most of the irregularities of the preglacial surface were filled in with older drift, so that, in contrast with the Rock River Hill Country, only gross features of the bedrock topography are reflected in the present landscape.

*Springfield Plain.*—The Springfield Plain includes the level portion of the Illinoian drift-sheet in central and south-central Illinois. It is distinguished mainly by its flatness and by shallow entrenchment of drainage as compared with the more sharply incised valleys of the Galesburg Plain. The southern boundary of the district, which coincides closely with a similar division made by Paul McClintock in 1929 (fig. 1, p. 28), is drawn along a line south of which the drift thins and bedrock topography becomes a controlling factor; the western boundary follows the edge of the Illinoian drift.

Although the greater part of the district is a flat till plain, the morainic features in the western part of the region are much more conspicuous than elsewhere on the Illinoian drift-sheet. They include the Jacksonville and Buffalo Hart moraines and an extensive area of ridged drift in the Kaskaskia drainage basin. The moraines are low and broad, but they are readily recognized because of their continuity and the associated kames and kame terraces. The Kaskaskia ridges lie just to the east of an interlobate area indicated by the moraines and include an irregular assemblage of gravelly ridges and hills with small intervening plains, some of which are old lake basins. A large proportion of the hills and ridges appear to be kames and crevasse-fillings related to stagnant ice conditions (Ball, 1940, pp. 951–970).

Drainage systems are well developed, and the district as a whole is in a late

youthful stage of dissection. The uplands are low with respect to the masterstreams, and the valleys are relatively shallow. Most of the principal streams have low gradients and occupy broad alluviated and terraced valleys; the secondary tributaries have wide V-shaped valleys; and the headwaters, flowing essentially on the till plain, have broad shallow valleys and low gradients.

The Illinoian drift is moderately thick and is underlain by older drift except in areas where the bedrock is close to the surface. Only the larger valleys and uplands of the bedrock surface are reflected in the present topography (fig. 3).

Along the southeast side of the Illinois Valley there is a belt of thick loess, with dune-contours characterizing the bluff-margin, but this body of loess thins rapidly to the southeast.

*Mount Vernon Hill Country.*—The Mount Vernon Hill Country comprises the southern portion of the Illinoian drift-sheet in Illinois and is characterized by mature topography of low relief with restricted upland prairies and broad alluviated valleys along the larger streams. Except for a southern extension of the Jacksonville moraine, glacial land forms are essentially absent. The southern and western boundaries of the district coincide closely with either the outer limits of glaciation or the outer margin of the Carbondale group of the Pennsylvanian system.

A relatively complete drainage system is present, and most streams have broad terraced valleys and low gradients. Natural drainage is good throughout the upland area, but the larger valley bottoms are poorly drained. Extensive aggraded lowlands along the Wabash drainage system to the east and the Big Muddy basin to the west are outstanding physiographic features.

The Illinoian drift in the district is thin, and deposits of underlying older drift are not known to be present except west of Sparta in Randolph County. The present land surface is primarily a bedrock surface of low relief, which is only slightly modified and subdued by a mantle of drift (fig. 3).

*Geologic history of the Till Plains section.*—Prior to glaciation the Till Plains section had a long and complex erosional history (Horberg, 1946, pp. 179–192). An extensive lowland—the central Illinois peneplain—was eroded in the weak Pennsylvanian rocks of the Illinois basin east of the Illinois River; it was bordered on the west and south by uplands, on which remnants of an older erosion surface are extensively preserved. Just prior to glaciation a system of deep bedrock valleys, many of which are occupied by present streams, were entrenched below the level of the central lowland. The gross features of the section as well as local features in the Rock River Hill Country and in the Mount Vernon Hill Country are determined to a large degree by this preglacial topography. The greater relief and higher elevations in the Rock River Hill Country and in the Galesburg Plain are determined by the preglacial uplands, whereas the low plains in the remaining districts reflect the central Illinois peneplain.

With the advent of glacial conditions and the approach of the Nebraskan glacier, there was probably a change from erosion to aggradation along major streams as the result of increased load and drainage derangements. The oldest deposits along the ancient Mississippi Valley (middle and lower Illinois) and its buried eastern fork, Mahomet Valley, appear to represent this stage. This was followed by the Nebraskan glacial invasion, which is known to have covered at

least a large part of the upland of western Illinois. There is no evidence that the early fills in the preglacial valleys were more than partially removed during the succeeding Aftonian interglacial stage. The Kansan glaciers, which followed, advanced from both the northeast (Labradorean center) and northwest (Keewatin center), probably in that order, and together covered most of the district except for the Rock River and Mount Vernon hill lands. Mahomet Valley and its tributaries in the central part of the section were largely buried and, because of the diversion of drainage, were not re-excavated during the ensuing Yarmouth interglacial stage (fig. 3) (Horberg, 1945, pp. 349–359).

With the advance of the Illinoian glacier from the Labradorean center, the ice attained its maximum extent in Illinois, and the entire Till Plains section was ice-covered. Except in the Rock River and Mount Vernon districts, where older drift is largely absent, the ice moved across a subdued land surface with fills of early drift and during retreat left behind a relatively smooth till plain. Following Sangamon erosion, which was not important except locally, the Wisconsin glaciers of the Tazewell stage covered the northeastern part of the state and formed the glacial landscape which is still so extensively preserved.

### DISSECTED TILL PLAINS SECTION

The Dissected Till Plains section in western Illinois is represented by a narrow isolated segment of the Kansan drift-sheet maturely dissected into an upland of high relief. The eastern boundary is determined by the Illinoian drift margin and the southern boundary by an arbitrary line, south of which the drift occurs as patches and is unimportant physiographically. From a regional standpoint, essentially the entire section lies west of the Mississippi River, and the eastern boundary of the section was drawn along the river by Fenneman. The refinement of the boundary here proposed is of only local significance.

The Kansan drift in the area is thin, and the topography closely reflects the ruggedness of the underlying bedrock upland. Valley-train terraces along the Missisippi Valley and thick loess deposits on the adjoining bluffs are features of secondary importance. The geomorphic history closely parallels that of the adjoining Ozark Plateaus and is discussed later.

### WISCONSIN DRIFTLESS SECTION

The Wisconsin Driftless section, which constitutes the final subdivision of the Central Lowland province in Illinois, is a submaturely dissected, low plateau bordering the outwash-filled valley of the upper Mississippi River. The eastern boundary follows the edge of the Illinoian drift and is unchanged from Fenneman's classification.

The upland is underlain by the Galena-Platteville dolomite and Maquoketa shale of Ordovician age and by outliers of Silurian dolomite. Flat upland areas, which coincide closely with the top of the Galena-Platteville dolomite, are considered remnants of the Lancaster peneplain.[1] A possible higher surface—the Dodgeville peneplain—may be represented by the crests of mounds and narrow ridges capped by Silurian dolomite. Benches clearly controlled by structure occur along the lower reaches of the principal valleys, where the Maquoketa shale has been stripped from the top of the Galena-Platteville dolomite.

The plateau is maturely to submature-

[1] See Horberg (1946) for bibliography and recent review of the problem.

ly dissected by a number of dendritic drainage systems tributary to the Mississippi. The Mississippi Valley has broad terraced bottom lands and precipitous walls. Most of the minor valleys are youthful, with narrow V-shaped valleys and some with incised meanders. There is considerable underground drainage through small caves and solution channels, but sinkholes and other karst features are not conspicuous. The canyon of Apple River is a prominent local feature resulting from glacial diversion of the former headwaters of Yellow River (Trowbridge and Shaw, 1916, pp. 95-99). As elsewhere, thick loess deposits mantle the bluffs of the Mississippi Valley and thin eastward.

The geomorphic history of the region is largely one of stream erosion and involves numerous uncertainties. The broad outlines, however, may be summarized as follows: (1) development of the Dodgeville surface, probably as a peneplain; (2) rejuvenation and erosional development of the Lancaster surface to a partial peneplain; (3) uplift and entrenchment of the Mississippi bedrock valley and some of its larger tributaries; (4) partial filling of the valley by glacial outwash with a maximum thickness of more than 300 feet; and (5) postglacial erosion.

### OZARK PLATEAUS PROVINCE

The Ozark Plateaus province forms a discontinuous upland along the southwest margin of the state and represents the eastern edge of an extensive upland in southern Missouri and northern Arkansas (fig. 2). It includes the driftless and thinly drift-covered cuestas on pre-Pennsylvanian rocks which are structurally and topographically a part of the Ozark dome. Two important modifications of Fenneman's classification are proposed: (1) the Salem Plateau section is expanded northward to include the partially drift-covered Mississippian cuesta in Randolph, Monroe, and St. Clair counties, which is clearly a part of the Ozark dome; (2) the Lincoln Hills section, first distinguished by E. M. Shepard (1907, pp. 8, 10-11) in Missouri and later by W. W. Rubey (1936) in Calhoun County, Illinois, is recognized as a new subdivision. Fenneman included both these areas in the Till Plains section of the Central Lowland province.

*Lincoln Hills section.*—The Lincoln Hills section includes the partially drift-covered dissected plateau above the junction of the Mississippi and Illinois rivers in western Illinois. It is part of a larger upland which, bisected by the Mississippi, lies partly in Missouri. The principal physiographic feature in Illinois is a maturely dissected central ridge, which forms the watershed between the Mississippi and the Illinois rivers throughout the length of the section. As previously noted, the northern boundary is arbitrary, and the eastern boundary follows the Illinoian drift border. The southern boundary with the Salem Plateau is drawn along the Cap au Grès flexure in southern Calhoun County.

The upland is determined by a subsidiary structure of the Ozark dome, the Lincoln fold, along which the more resistant pre-Pennsylvanian limestones and dolomites crop out. In Illinois the plateau is largely underlain by Osage limestones, of which the Burlington limestone is most important physiographically; and the boundaries coincide quite closely with the Mississippian-Pennsylvanian contact. The southern part of the section is driftless except for loess deposits and a single high-channel filling of outwash gravel, presumably Kansan. It has long been known as the Calhoun

County driftless area. Patchy remnants of Kansan drift are preserved in the northern part of the section.

The plateau surface is rugged and broken by closely spaced valleys and ridges. Restricted areas of flattish to gently rolling upland representing the Calhoun peneplain (Rubey, 1936) are present along the crest of the ridge. The valleys of the Mississippi and Illinois rivers are broad, deeply alluviated, terraced, and have precipitous walls. Most of the minor valleys are narrow, V-shaped, and steeply graded.

*Salem Plateau section.*—The Salem Plateau comprises the major part of the Ozark dome in southern Missouri, but only two small segments, isolated by the Mississippi River, are present in southwestern Illinois. Both segments are maturely dissected, partially truncated cuestas, dominated by a single central ridge. The northern segment is covered by thin Illinoian drift, but the southern segment lies south of the glacial boundary. In the northern segment an arbitrary boundary with the Shawnee Hills is drawn where the sandstones and conglomerates forming the lower Pennsylvanian escarpment give way to finer sediments and the escarpment dies out, the east margin closely follows the overlapping edge of Pennsylvanian strata, and the northern boundary coincides with the Cap au Grès flexure. The southern segment is delimited from the Shawnee Hills to the east along the contact between Carboniferous and older rocks and follows Fenneman's boundary.

The northern segment is developed on Mississippian strata and lies on the back slope of the Meramec-Osage cuesta, which flanks the Ozark uplift on the north and east. It is underlain by Meramec limestones on the west and north and by Chester strata on the southeast. The plateau is submaturely dissected; and gently rolling summit areas, considered remnants of the Ozark peneplain, occur along the central ridge. Because of the drift mantle, the topography does not appear as rugged as the Lincoln Hills or the Salem Plateau sections. Karst features, developed primarily on the St. Louis limestone, are present at many places within the area. The central ridge forms the watershed for tributary drainage, but the Mississippi and Kaskaskia rivers cross the ridge without regard to structure. The valleys of these two major streams have broad alluvial flats and steep walls, whereas most of the tributaries are youthful.

The south unglaciated segment of the Salem Plateau in Illinois is underlain largely by a thick succession of deeply weathered Devonian chert and cherty limestone formations which on the south are overlapped by Coastal Plain sediments. Structurally, the area is clearly a part of the Ozark dome, but it is complicated by a zone of folds and faults trending north-south and northwest-southeast. A clearly defined physiographic boundary separates the plateau from the Shawnee Hills to the east and north, the contrast being marked by more rugged hills, closer drainage texture, absence of structural control, and higher elevations in the plateau section. Most of the plateau is maturely dissected by intricate dendritic drainage, although small remnants of a flat upland surface representing the Ozark peneplain are preserved throughout the region. The northern part of the segment is drained by streams which head in the Shawnee Hills and flow westward across the plateau into the Mississippi River, whereas in the southern part a central divide separates the Mississippi and Cache Valley drainage. In contrast to other parts of the Ozark

Plateau in Illinois, most of the larger tributary valleys, as well as the Mississippi Valley, are deeply alluviated, and only the secondary tributaries are youthful.

*Geomorphic history.*—The Ozark Plateaus are essentially a preglacial land surface whose erosional history has continued during the glacial period. The oldest landscape features in the province are isolated summit areas, over 800 feet above sea-level, which may be peneplain remnants correlative with the Dodgeville peneplain of the Wisconsin Driftless section and the Buzzard's Point plain (Salisbury, in Weller, Butts, Currier, and Salisbury, 1920, pp. 47–52) of the Shawnee Hills. An alternative interpretation is that they are simply monadnocks on the lower Ozark peneplain. In either case an extensive surface, called the "Calhoun peneplain" in the Lincoln Hills section and the "Ozark peneplain" in the Salem Plateau section, was developed below the level of these isolated remnants and is responsible for the general accordance of summit levels found throughout the plateau at elevations about 700 feet above sea-level. Near the southern margin of the plateau the Ozark peneplain is believed to transect Wilcox (Eocene) strata and therefore to have been completed sometime during the Tertiary (Flint, 1941, pp. 634–636). The weathering and leaching of the Devonian formations in the southern part of the Salem plateau to depths of about 400 feet is of unusual interest and has been ascribed to prolonged alteration under peneplain conditions (Weller, 1944, pp. 101–102). Following completion of the peneplain, and probably prior to erosion of the Central Illinois peneplain, "Lafayette"-type gravels were spread over its surface. It appears likely that the major preglacial drainage lines were determined at this time and that, with uplift of the peneplain, streams occupying the Mississippi, Illinois, and Kaskaskia valleys became incised without regard to structure. There is no clear evidence of the succeeding central Illinois peneplain and Havana strath cycles in the region, probably because the weaker formations on which they are elsewhere developed are absent. During the glacial period the preglacial topography was modified by alluviation of the major valleys and by deposition of loess on the uplands.

### INTERIOR LOW PLATEAUS PROVINCE

*Shawnee Hills section.*—The Interior Plateaus in southern Illinois are represented by the western part of the Shawnee Hills section[2] and include a complex dissected upland, underlain by Mississippian and Pennsylvanian strata of varied lithology. It is, in the main, the area generally referred to popularly as the "Illinois Ozarks." The northern margin is drawn along a marked topographic boundary which lies along the inner flank of the lower Pennsylvanian (Caseyville) cuesta just within the Illinoian glacial drift boundary, and the southern boundary follows the northern edge of the overlapping Coastal Plain sediments. These are essentially Fenneman's boundaries, the only important modification being a northwestward extension of the section to include the thinly drift-covered Pennsylvanian cuesta in Jackson and Randolph counties.

The section is situated along the southern rim of the Illinois basin, so that the lower Pennsylvanian cuesta com-

[2] The section was originally distinguished by R. F. Flint (1928, pp. 451–457) and named the "Shawnee Hill Section. Fenneman (1938, p. 435) used the name "Shawnee section." The suggested usage of the term "Shawnee Hills section," in the present report is proposed purely for descriptive reasons.

prises the northern part of the region and a dissected plateau underlain largely by Chester (Mississippian) formation comprises the southern part. This regional structure is complicated by faulting and folding, which to a varying degree involved a large part of the area.

The Pennsylvanian cuesta forms a continuous ridge and watershed, extending completely across the state. In most places the ridge is maturely dissected by youthful valleys, but remnants of flat upland are locally preserved on narrow ridge crests throughout the length of the escarpment.

The plateau on Mississippian rocks to the south is maturely dissected, and the larger valleys are alluviated. There are numerous minor escarpments, structural benches, fault-line scarps, and subsequent valleys which reflect local structure and the varied lithology of the bedrock. Only small patches of flat upland are present. Karst features in the St. Louis limestone are present near Cave in Rock, Hardin County, and in south-central Union County.

*Geomorphic history.*—The erosional history of the region is similar to that of the Ozark Plateaus previously outlined. Remnants of the Ozark peneplain appear to be extensive along the Pennsylvanian escarpment, and local higher summits, especially to the east, may represent an older (Buzzard's Point plain) cycle (Salisbury, in Weller, Butts, Currier and Salisbury, 1920, pp. 47–52). Lower surfaces on the Mississippian rocks, 500–550 feet and 600–650 feet in elevation, are of uncertain origin. Remnants of "Lafayette"-type gravels are found both on the escarpment and at lower elevations south to the Ohio River. A deep weathered zone on the gravels overlain by loess indicates that a long period of stable conditions followed their deposition and that the major period of valley-cutting occurred late in the Tertiary (Weller, 1940, p. 45). Loess deposition and valley alluviation were the principal events during the glacial period.

#### COASTAL PLAIN PROVINCE

The Coastal Plain in Illinois includes the southern tip of the state and is underlain by unconsolidated Cretaceous and Tertiary sediments, which overlap the older Paleozoic rocks to the north. Three physiographic subdivisions are recognized: (1) and (2) the coextensive alluvial plain of the Cache and Mississippi valleys and (3) the Cretaceous hills between the Cache Valley and the Ohio River. The alluvial plains are characterized by terraces and recent floodplain features. The Cretaceous hills are maturely eroded into a low upland of gently sloping knolls and ridges. Outwash and alluvium extend far up tributary valleys, so that the upland is partially buried and certain segments are essentially isolated.

The earliest events in the geomorphic history of the region are indicated by remnants of "Lafayette"-type gravels which occur in the Cretaceous hills. The erosion surface at their base is evidence of a long period of denudation, during which the Coastal Plain deposits were lowered and stripped back. This was followed by deposition of the gravels, their weathering under stable conditions, establishment of major drainage lines, and final dissection of the bedrock topography. Prior to glaciation, Cache Valley was occupied by the Ohio River and the present Ohio Valley was occupied by the Cumberland and Tennessee rivers. During Illinoian or possibly Wisconsin time the valleys were aggraded to the level of the divide between the Ohio and the Cumberland rivers at Bay City in south-

ern Pope County, and the present lower course of the Ohio was opened. Both courses were kept open during subsequent stages, so that flood waters still pass through Cache Valley, and it was only in relatively recent time that the southern channel became the permanent course of the river.

## REFERENCES CITED


Ball, J. R. (1940) Elongate drift hills of southern Illinois: Geol. Soc. America Bull. 51.

Bretz, J H. (1939) Geology of the Chicago region: Illinois Geol. Survey Bull. 65, pt. 1.

Ekblaw, G. E., and Athy, L. F. (1925) Glacial Kankakee torrent in northeastern Illinois: Geol. Soc. America Bull. 36.

Fenneman, N. M. (1914) Physiographic boundaries within the United States: Annals Assoc. Am. Geog., vol. 4.

——— (1928) Physiographic divisions of the United States: *ibid.*, vol. 18.

——— (1931) Physiography of western United States, New York, McGraw-Hill Book Co.

——— (1938) Physiography of eastern United States, New York, McGraw-Hill Book Co.

Flint, R. F. (1928) Natural boundaries in the Interior Low Plateau physiographic province: Jour. Geology, vol. 36.

——— (1941) Ozark segment of Mississippi River: Jour. Geology, vol. 49.

Fryxell, F. M. (1927) The physiography of the region of Chicago, Chicago, University of Chicago Press.

Gannett, Henry (1892) The average elevation of the United States: U.S. Geol. Survey 13th Ann. Rept.

Hershey, O. H. (1893) Pleistocene rock gorges of northwestern Illinois: Am. Geologist, vol. 12.

Horberg, Leland (1945) A major buried valley in east-central Illinois and its regional relationships. Jour. Geology, vol. 53.

——— (1946) Preglacial erosional surfaces in Illinois: Jour. Geology, vol. 54. Reprinted as Illinois Geol. Survey Rept. Inv. 118.

Leighton, M. M. (1923) The differentiation of the drift sheets of northwestern Illinois: Jour. Geology, vol. 31.

McClintock, Paul (1929) Physiographic divisions of the area covered by the Illinoian drift sheet in southern Illinois: Illinois Geol. Survey Rept. Inv. 19. pt. 1.

Poggi, E. M. (1934) The Prairie province of Illinois: Univ. Illinois Studies in Social Sci., vol. 19.

Rubey, W. W. (1936) Geology and mineral resources of the Hardin-Brussels quadrangles, unpublished manuscript of Illinois Geol. Survey.

Shepard, E. M. (1907) Underground waters of Missouri: U.S. Geol. Survey Water Supply Paper 195.

Trowbridge, A. C., and Shaw, E. W. (1916) Geology and geography of the Galena and Elizabeth quadrangles: Illinois Geol. Survey Bull. 26.

Weller, J. M. (1940) Geology and oil possibilities of extreme southern Illinois: Illinois Geol. Survey Rept. Inv. 71.

——— (1944) Devonian system in southern Illinois: Illinois Geol. Survey Bull. 68A: Devonian symposium.

Weller, Stuart; Butts, Charles; Currier, L. W.; and Salisbury, R. D. (1920) Geology of Hardin County: Illinois Geol. Survey Bull. 41.

William, H. B., and Payne, J. N. (1942) Geology and mineral resources of the Marseilles, Ottawa, and Streator quadrangles: Illinois Geol. Survey Bull. 66.

# THE FORMATION OF BEACH CUSPS

Ph. H. KUENEN
Geological Institute, Groningen, Netherlands

ABSTRACT

In former attempts to explain the development of beach cusps, stress has been laid on the erosion of the beach. It is argued in this paper that concomitant deposition on the horns is at least equally important. Refraction of the swash in building the cusps in emphasized, and an attempt is made to explain the rhythmic, roughly equidimensional nature of cusps.

## INTRODUCTION

Beaches sometimes show regularly spaced, crescentic accumulations of materials, ranging from sand to cobbles. These formations are now generally termed "beach cusps." The projecting parts are more or less triangular, with a rounded apex extending into the water and with curved bays between the promontories. The basal, or landward, parts of the triangle may be horizontal, with the apex sloping down the beach. In many cases the cusps are of coarser material than that of the remainder of the beach, and they either merge gradually into the latter or are set off sharply from them. The height is usually very slight but may also attain more than a meter, and the distance between cusps ranges from a few centimeters to several dozens of meters. The apex may protrude a few centimeters to several meters, and the relative depths of the bays are also variable. In front of the bays the foreshore is built out under water in a delta shape (fig. 1).

Although cusps may develop on many types of beach, they are most common on slightly concave stretches of the coast. Observation has shown that they may be formed or destroyed in a few hours, small ones even more quickly. Each change in size of the waves results in the development of a new series of cusps to fit the altered conditions. According to Douglas Johnson (1919), the interspace is roughly doubled for a doubling of the wave height.

O. F. Evans (1938) made observations on lake shores and divided the observed cusps into five classes. Two of these occur as individual accumulations, not as a rhythmic series. His "giant cusps" are formed by erosion and deposition during storms. Very small cusps a few inches apart are the result of a gentle onshore "slop" of a "dead sea" and are really a series of rill marks. None of these will be considered in this paper, but the giant cusps are practically the same as his fifth class. These "ideal cusps," which usually occur in series, are similar to those on sea beaches. Evans showed by measurement that the interspace generally varies about 50 per cent, occasionally over 100 per cent. Although these variations are appreciable, they confirm the conclusion that the phenomenon is essentially of a rhythmic nature. Any attempt at explanation must show that the process is not arbitrary in its action but that it contains elements regulating the cusp-interspace.

## ORIGIN OF BEACH CUSPS

The origin of beach cusps is not fully understood. Escher (1937) was able to produce cusps experimentally by the action of ordinary waves, combined with standing waves, the latter at right angles

to the model beach. This explanation cannot be applied to beach cusps until proof is given of the occurrence in nature of strongly developed standing waves with wave lengths comparable to the size of cusps. No records of such standing waves are known to the present author, and it appears highly doubtful whether they could develop over distances of hundreds of meters. According to Johnson (1919), erosion of the beach by the swash plays an important part. Slight irregular depressions undergo erosion by the waves because more water rushes in and out than up and down the adjoining even slopes. The larger a depression becomes, the more strongly it will grow. These depressions are enlarged until they together occupy the whole beach, with narrow tongues between. The coarse material tends to be thrown up the beach and onto these projecting cusps. Johnson (1919) and Timmermans (1935) succeeded in producing small cusps experimentally in a tank with a wave-producing apparatus. According to Evans, these were not true cusps, but of the nature of his "very small cusps."

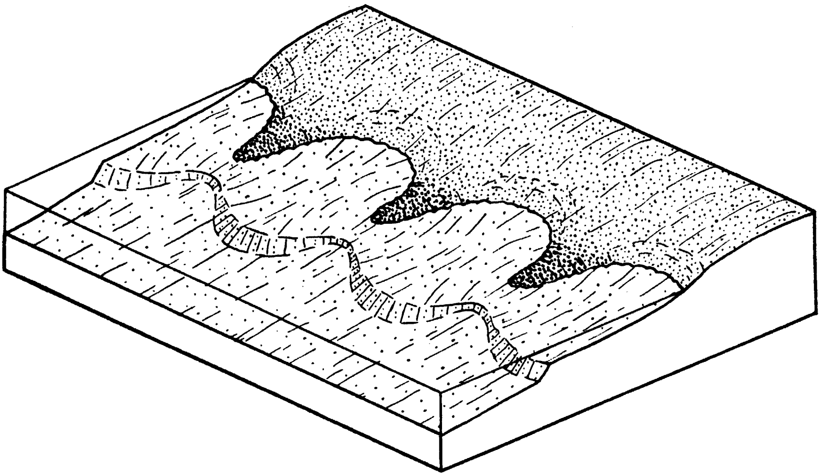


FIG. 1.—Block diagram of beach cusps, showing the horns (in this case consisting of coarse material) and the submarine deltas corresponding to the embayments. (Mainly after Timmermans.)

This explanation is certainly more satisfactory than many other suggestions that have been offered. But several points remain obscure. In the first place, the beach as a whole is not eroded during the development of cusps, but in Johnson's theory only erosion is mentioned. Evidently, accumulation goes on concomitantly with erosion, and the horns must represent not merely buttresses left standing by the erosive action in the bays but prograded areas where most of the eroded material comes to rest. The fact that the cusps generally consist of material differing from the remainder of the beach is further evidence of the selective transportation from bay to horn and the outbuilding of the latter.

In the second place, it is not evident why a stable condition is attained, once the cusps fitting the waves are developed. Johnson says (1919, p. 483): ". . . . enlargement will continue only so long as the impulse toward growth im-

posed on the more favored channels is sufficiently great to overcome the tendency of their neighbors to enlarge. Equilibrium will be established when adjacent channels are approximately the same size, and at the same time of a size appropriate to the volumes of water traversing them." But, as further erosion would result in a larger volume of water running in and out of an embayment, one may ask what is the meaning of the size being "appropriate to the volume of water." Neither is it clear why the growth of its neighbors should impede the enlargement of a bay.

As soon as it is admitted, however, that the horns are built out by materials washed from the bays, it is obvious that there must be a close relation between the development of adjoining bays. Wherever one is enlarged, the concomitant growth of its horns must tend to obliterate its neighbors. In the sentence quoted from Johnson, we should make a small alteration as follows: "Enlargement of the bays *and cusps* will continue only so long as the impulse towards growth imposed on the more favored channels is greater than with its neighbors." It then remains only to be shown that, starting from small irregularities, the larger ones are, first, at an advantage but that, with increasing size, this impulse toward growth must eventually disappear, so that the less developed bays will then tend to enlarge at the cost of the larger, overgrown channels. Only in this manner can an evenly spaced series of cusps be accounted for.

Before attempting to show that this relation actually exists, the process by which the horns are prograded must first be found. The present writer has observed that *refraction of the swash*, as it sweeps into the embayments and fans out onto the sides of the protruding horns, is also an essential element in the production of the cusps. This causes pebbles to be washed sidewise out of the troughs onto the horns.

The backwash is somewhat concentrated toward the horn by the same process. The sidewise component of the uprush is directed toward the cusp from both sides (pl. 1, *A*); and, because we are dealing with swash and not with oscillation waves, an actual piling-up of water over the area of the horn is brought about. When the water begins to flow back under the influence of gravity ($a$–$f$ in fig. 2), it must trend somewhat in the direction perpendicular to the shore line ($a'$–$f'$) in consequence of the accumulation of water over the area *DEF*.

Hence a particle of water will follow a curved course over the cusp, somewhat similar to the movement during normal beach drifting. This curved path, directed toward the apex of the cusp, carries the pebbles out toward the tip of the horn. But, because waves tend to throw pebbles high up onto the beach and carry the sand fraction away, the pebbles are deposited on the cusps, and the sand is rolled back into the bay. The next rush of water will bring part of the sand onto the beach of the embayment. The remainder will be dropped on the delta scallops in front of the bays.

Another consequence of the refraction of the swash is that the strengthened backwash over the cusps impedes the next swash on the horn and thus helps to concentrate erosion in the embayment. A complication arises because the swash is impeded by the backwash more strongly in the bays and on the horns alternately. The writer has not been able to make sufficient observations under varying conditions to ascertain whether or not this alternate action is the rule and forms an essential process in the develop-

PLATE 1

A

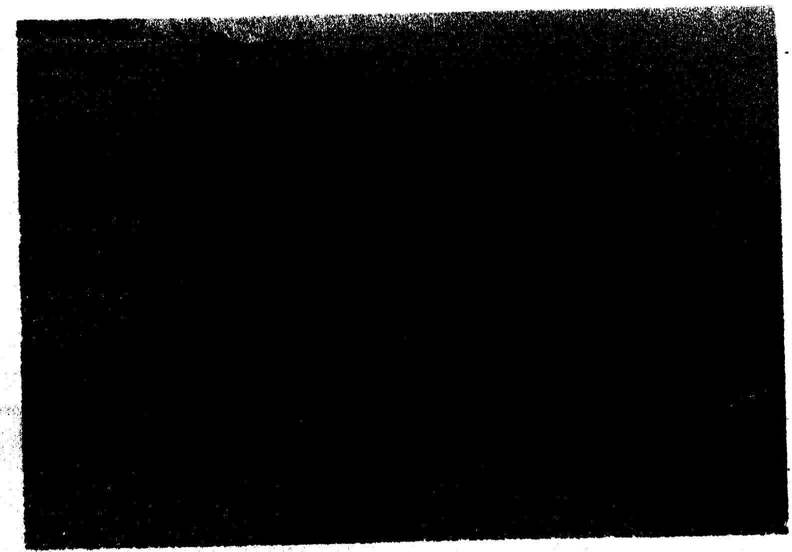
*A*. The same as Pl. 1, *B*, at advanced stage in uprush of the swash. Note the marked refraction of the swash and advance of the water from both sides on the cusps.

B

*B*, Beach cusps, Alum Bay, Isle of Wight. Apex to apex is 8 meters.

A

*A*, Waves playing on cusps in Sandown Bay, Isle of Wight, at low tide. Concentration of backwash at intervals corresponding to the rhythm of the intercusp space.



B

*B*, The same as *A* at slightly later stage in the uprush of the swash.

ment of cusps. The photograph (pl. 2, *A*) was made during low tide, when the waves were playing only at the lower end of the cusps. It shows the concentration of the backwash at intervals corresponding to the rhythm of the intercusp space. Plate 2, *B*, shows a slightly later stage in the uprush of the swash.

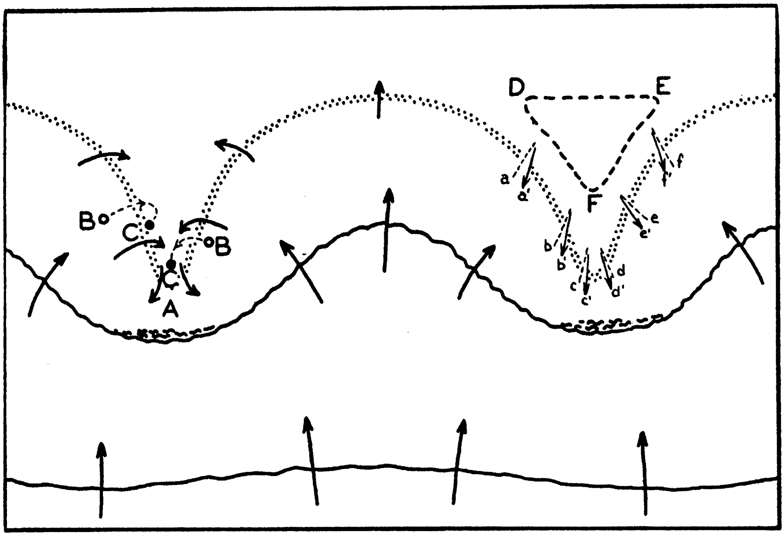



FIG. 2.—Diagram showing the diffraction of the swash. Pebbles are rolled from *B* to *C*. At *A* the powerful backwash of the foregoing wave impedes the swash and increases the diffraction. *a*–*f* = slope, *a′*–*f′* = direction of backwash owing to the piling-up of water over the area *DEF*.

Although Evans (1938) speaks of the "parabolic path" of the water and refraction around the cusp apex, he does not show in what manner this influences the development of cusps.

Evans came to the same conclusion on the origin of cusps as did Jefferson, who also studied lake-beach cusps. Both authors were able to show that on lake shores the breaching of a beach ridge or the edge of a low-cut bank is essential to the formation of cusps. On sea shores with tidal movements this is apparently not a necessary condition. Not only are cusps frequently found where no ridge appears to exist, but they may be observed in two sets, one directly below the other, where there is no space for a ridge in between (see Johnson, fig. 142). Moreover, Gellert (1937) saw cusps forming during a storm on the German North Sea coast, where no ridge existed. As I have made no observations on lake shores, I will deal only with the case of shores with tides.

Here a certain amount of erosion is probably also essential to the formation of cusps but not to the breaching of a ridge. Thus Timmermans (1935) found that on the Dutch coast and in his experiments cusps developed only where the beach was steeper than normal and erosion took place.

### SPACING OF CUSPS

We must now return to the question of even spacing. It is obvious that the degree of refraction and also its conse-

quences are closely related to the size of the waves and the slope of the beach. A small cusp cannot influence the shape of a powerful wave, and the horns will be attacked and washed away. Small waves breaking on large cusps are spaced too closely together so that there is no time for the backwash of the first to run out of the embayment before the next one comes rolling in. Moreover, they carry insufficient water to influence large cusps. Smaller irregularities will then be picked out and developed separately into a new set of narrowly spaced cusps, better adjusted to the small volume of the waves.

This rather vague relation between the size of waves and cusps can be worked out in greater detail. In the first place, it is evident that a certain size of wave cannot erode the bays beyond a certain depth of water because the volume of water passing in and out is constant, and, consequently, with increasing depth the current becomes too sluggish to move the sediment. The distance to which a bay can be eroded into the beach is thereby also limited, because, with a limit set to the depth, the outward slope of the bottom becomes so slight that gravity is unable to cause sufficient current for dragging sediment outward. In the second place, broadening of the bay delivers more material for building out the cusps. This, in turn, must cause prograding of the embayment floor and consequent weakening of the erosion in the bay. In the third place, outward growth of the horns must result in stronger attack by the swash, because the protecting influence of the backwash will decrease. This means that the material carried out toward the apex is brought back into the channel. It then moves back up the bay to the beach.

Summarizing the deductions given above leads us to the following explanation of cusp formation. On a smooth beach a regular train of waves will cause a succession of swash flows. These flows encounter slight depressions and start to erode them, while the backwash carries some sediment out of the embayments to build deltas opposite them. As long as the depth of water in a bay is so small that the water passing in and out is able to carry the sediment along, the enlargement will continue both horizontally and vertically. As the channel becomes deeper and its side slopes steeper, the breadth also tends to increase. But, when the depth in the outer portion of the bay approaches a certain limit, erosion gradually slackens. Refraction of the swash results in transport of material toward the sides and causes prograding at these points. The coarse material tends to be pushed back up the beach of the embayment and out along the developing cusps. Growth of the bays and prograding of the cusps must gradually decrease when the maximum depth in the central area of the bay has been attained.

In the meantime, adjacent channels have been subject to the same process. Where these are so close together that the two natural spheres of growth overlap, a rivalry develops. This should tend to push the bays farther apart, because the material eroded from the one and dumped onto the intervening cusp must tend to encumber the growth of the other. Erosion in the latter will be shifted slightly to the opposite side and thus cause the entire bay to move away.

Where two neighboring bays are farther apart, their adjacent cusps will not coalesce and will leave a small space between. In this space refraction begins, and a new bay will start to form.

As long as the maximum depth of water in a bay has not been attained, the

tendency to enlarge will be greater than in an adjacent, shallower indenture. The larger one will grow at the expense of the smaller one. This relation is reversed when the depth approaches its maximum value. Then a smaller bay is more powerfully eroded and tends to encroach on its overgrown neighbors. In this manner a balance will be reached when all bays have attained the maximum depth for the size of waves playing on them. They will push each other aside or obliterate some until this stage is attained and a regular rhythmic pattern has been evolved (fig. 3, pl. 1*B*).

The writer believes the process here pictured accounts for the main phenomena of cuspate beaches: the regularity of the pattern, the growth in size with increasing wave dimensions, and the accumulation of coarse material in the horns. But, although a step forward may have been made by this working hypothesis in explaining these fascinating features of the shore, several problems still await solution. The differences in size and shape, slope and grain size, are evidently interrelated, but in what manner is not known.

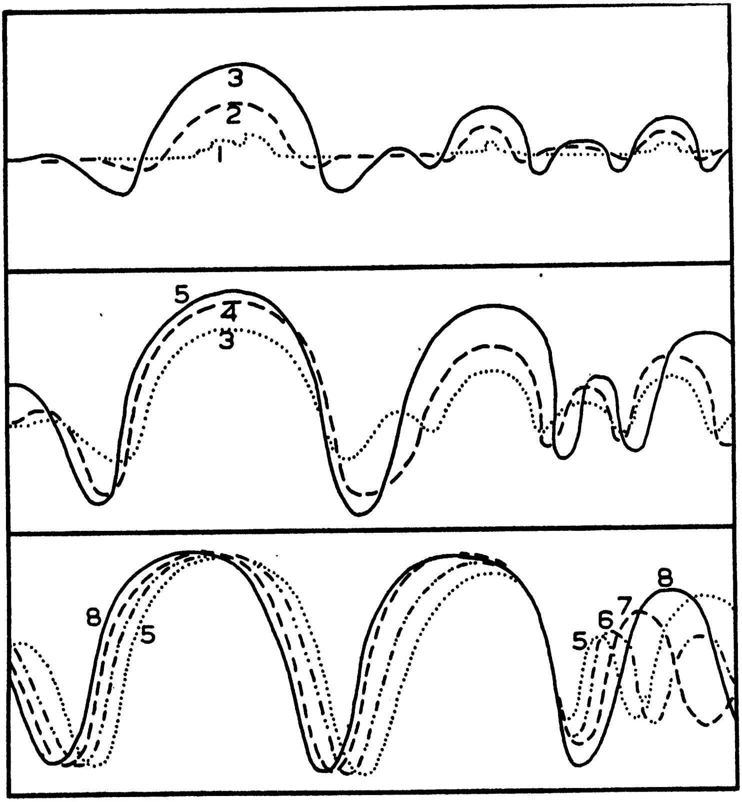



FIG. 3.—Successive stages in the development of cusps from irregular indentures of the beach to an almost regular, rhythmic pattern.

Neither has it been made clear why

cusps are of only exceptional occurrence. Johnson (1919) and Timmermans (1935) both believed that waves directed straight onshore are most effective in building cusps. Thus the latter found cusps on the Dutch coast only during periods of offshore winds when regular small waves advance straight toward the shore. This may explain why bays are favorable sites, because oblique waves cannot form so easily (Timmermans, 1935, p. 364). But, on the other hand, onshore waves may play for a long time on a beach without producing cusps. Moreover, several investigators have shown that cusps may be formed by oblique winds also.

The influence of tidal movements is another aspect in need of further study. Timmermans (1935) believed that the cusps are first formed during rising tide and then lengthened seaward during the falling of the water. F. P. Shepard maintains (personal communication) that the tides are important. Their action is evidently not necessary, because it has already been pointed out that cusps are formed on lake shores, where they are related to the breaching of a ridge. A special case in which erosion may be the main cause is that of boulder cusps, as described by Butler (1937, pp. 442–453).

Finally, there is an almost total lack of quantitative data and the flow of the water should be studied in much greater detail.

## REFERENCES CITED


Butler, J. W. (1937) Boulder beach cusps, Lake Olga, Quebec: Am. Jour. Sci., vol. 33, pp. 442–453.

Escher, B. G. (1937) Experiments on the formation of beach cusps: Leidsche Geol. Mededl., vol. 9, pp. 79–104.

Evans, O. F. (1938) Classification and origin of beach cusps: Jour. Geology, vol. 46, pp. 615–627.

——— (1945) Further observations on the origin of beach cusps: *ibid.*, vol. 53, pp. 403–404.

Gellert, J. F. (1937) Strandhörner bei Duhnen-Cuxhaven: Senckenbergiana, vol. 19, pp. 7–12.

Johnson, D. W. (1919) Shore processes and shoreline development, pp. 457–486, New York, John Wiley & Sons, Inc.

Krejci-Graf, K. (1933) Beobachtungen am Tropenstrand: Senckenbergiana, vol. 17, pp. 53–55.

Timmermans, P. D. (1935) Proeven over de invloed van golven op een strand (English summary): Leidsche Geol. Mededl., vol. 6, pp. 231–386.

# THE DISTRIBUTION OF OXYGEN IN THE LITHOSPHERE

TOM. F. W. BARTH
University of Chicago

ABSTRACT

Oxygen, which makes up more than 90 per cent by volume of the total lithosphere, shows the highest concentration in the outer shell. The regular decrease with depth represents an approximation to thermodynamic equilibrium. When highly oxidized surface rocks are brought down to great depths, oxygen will be squeezed out of the mineral lattices and returned to the surface. Therefore, the deeper parts of our globe cannot become oxidized.

## THE VOLUME RELATIONS OF OXYGEN IN THE LITHOSPHERE

The building bricks of the crystals are atoms (or ions); and in the crystalline edifice, as in other constructions, the largest bricks are the most important. In the late twenties it was recognized from X-ray studies that oxygen is among the very largest constituent ions of the rock-forming minerals; and it was appreciated that the oxygen ions, which had been generally neglected by chemists, were of fundamental importance in the physical chemistry of minerals.

The dominant role of oxygen in the lithosphere is demonstrated in table 1.

The igneous rocks and the whole lithosphere are to be regarded as essentially a packing of oxygen ions. The accumulation of this huge volume of oxygen is made possible through the cations, which occupy the interstices and, with their electrical charges, keep the whole structure together, although their volume is comparatively insignificant.

It is interesting to note that the lithosphere contains more oxygen (in terms of atom, weight, and volume percentages) than does the atmosphere. Thus the lithosphere appropriately might be called the "oxysphere" (Goldschmidt, 1928).

Estimates of the chemical composition of the interior of the earth are uncertain—indeed, they are based on extrapolations into the unknown. According to our best information, the outer shell, the so-called "crust," is made up of the following concentric layers which inperceptibly merge into one another:

1. Sediments, which are partly penetrated by
2. "Sial," which extends to a depth of approximately 25 km. Its composition corresponds to the computed average for all igneous rocks, with granitic rocks at the top and gradually changing downward into
3. "Sialma," which occupies the depth-range 20–70 km. and whose average composition is that of a plateau basalt.
4. At still greater depth, heavy basic silicates prevail. This region has been referred to as the "eclogite or peridotite shell." Typically, olivine is present.

Average figures for the oxygen content of the several shells in the earth are listed in table 2. The results are shown graphically in figure 1.

## CONDITIONS OF THERMODYNAMIC EQUILIBRIUM IN A FIELD OF GRAVITATION

I shall endeavor to show that the small but consistent drop in oxygen with depth as demonstrated by these data represents an approximation to thermodynamic equilibrium.

In the chemical laboratories we are wont to consider equal vapor tensions in all phases as necessary and sufficient for equilibrium. In a gravitational field this is different, a fact which has been pointed out and emphasized by Ramberg (1944*a*; 1944*b*, pp. 98–111; 1945, pp. 307–326; 1946, pp. 13–29). Various objections, mostly due to misunderstandings, have been raised to Ramberg's principle. To clarify the issue, the following restatement of the principle is offered.

TABLE 1

CHEMICAL COMPOSITION OF TYPICAL ROCKS, EXPRESSED IN VOLUME PERCENTAGES

| | Average Basalt* | Average Igneous Rock† | "Ichor" Granite‡ | Ionic Radii |
|---|---|---|---|---|
| O | 91.11 | 91.83 | 92.12 | 1.32 |
| Si | 0.70 | 0.83 | 0.92 | 0.39 |
| Ti | 0.12 | 0.05 | 0.02 | 0.64 |
| Al | 0.74 | 0.79 | 0.76 | 0.57 |
| Fe | 1.47 | 0.58 | 0.21 | {0.67, 0.82 |
| Mg | 1.09 | 0.58 | 0.09 | 0.78 |
| Ca | 2.78 | 1.50 | 0.45 | 1.06 |
| Na | 1.28 | 1.64 | 1.75 | 0.98 |
| K | 0.70 | 2.19 | 3.68 | 1.33 |
| Total | 99.99 | 99.99 | 100.00 | ........ |

* Average of Plateau basalt, 50 flows. Recalculated from Daly (1927).

† Recalculated from Clarke and Washington (1924).

‡ Typical pre-Cambrian "petroblastic" granite from Birkeland, southern Norway (Barth, 1948).

All three analyses are recalculated *on a water-free base.* Manganese is included in iron.

TABLE 2

OXYGEN IN PERCENTAGE BY VOLUME

| | |
|---|---|
| Air | 20.95 |
| Ocean | 96.87 |
| (Quartz | 98.73) |
| "Ichor" granite | 92.12 |
| Average igneous rocks | 91.83 |
| Average basalt | 91.11 |
| (Olivine* | 90.00) |

* 90 $Mg_2SiO_4$, 10 $Fe_2SiO_4$.

We will consider a vertical section through a homogeneous part of the lithosphere: The chemical potential of a species depends here on its position in the vertical section. The differential equation relating the chemical potential, $\mu$, of a chemical species with its position in the gravitational field is

$$\left(\frac{\partial \mu}{\partial x}\right)_{P,\,T} = M\,,$$

in which $M$ is the molecular weight of the species and $x$ is the vertical distance.

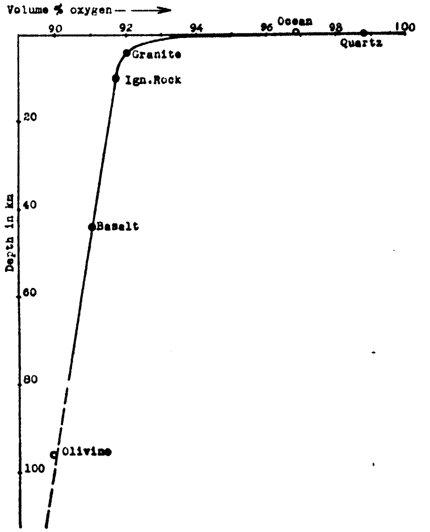



FIG. 1.—The volume percentages of oxygen at various depths.

Consequently, at the depth, $x$, the chemical potential is

$$\mu_x = \mu_0 - M \cdot X\,. \qquad (1)$$

Thus, the lower the position, the lower the chemical potential. In order to restore equilibrium, the chemical potential must be raised correspondingly. The raise may be effected by the superincumbent load, $P$:

$$\mu_x = \mu_0 + PV\,, \qquad (2$$

in which $V$ is the fictive volume of the species. Obviously, the term $M \cdot X$ in

equation (1) corresponds to the loss of potential mechanical energy per mol over a distance $X$; but it also corresponds to the weight of the superincumbent load if the chemical species with which we are dealing is overlain by itself.

The oxygen ions in the mineral lattices of the lower part of the lithosphere show a chemical potential which is higher than that required to restore equilibrium; for the overlying rocks exhibit a much higher average molecular weight than that of oxygen; this means lack of equilibrium. A manifestation of the disturbed equilibrium and of the high $\mu$-values thus generated at great depths is a high vapor tension and a consequent tendency of the oxygen ions to escape and migrate upward to places of lower chemical potential.

Stated in simple language, it means that, at high pressures, oxygen is squeezed out of the crystalline lattices, whereby compounds containing less oxygen are formed. Thus is established a vertical oxygen gradient in the lithosphere.

Ramberg (1946) has pointed out that water-bearing minerals are not stable at greater depths but give off water, which migrates upward. Nor are high oxides of iron stable; at greater depths they dissociate, with formation of oxygen, which also migrates upward.

### THE DEPTH RELATIONS OF $Fe_2O_3$, $Fe_3O_4$ FeO, AND IRON

As an example of this process, we can calculate the depth at which oxygen is squeezed out of the iron oxides.

Let us start with a vertical section of the lithosphere containing pure hematite, $Fe_2O_3$. For the sake of simplicity in this theoretical discussion we shall disregard the upper zone of weathering and oxidation; we postulate an ideal surface of demarcation separating this zone from deeper-lying material that is not in physical contact with, or exposed to, the oxygen of the surface waters and of the atmosphere. Likewise, for the sake of simplicity, we assume that the whole section is of uniform temperature.

The chemical potentials of the oxygen ions in the hematite lattice increase with the superincumbent load in accordance with equation (2). Obviously, we have

$$P = X \cdot d_h$$

and

$$V = \frac{M}{d_0} \text{ (by definition)},$$

where $d_h$ and $d_o$ are the specific gravities of hematite and oxygen respectively (see table 4).

Substituting in equation (2), we obtain as an expression of the actual increase of the chemical potential of oxygen with depth the following equation:

$$\mu_x = \mu_0 + M \cdot X \cdot \frac{d_h}{d_0}. \tag{3}$$

This expression should be compared with equation (1). It is seen that the factor $d_h/d_o$ is a measure of the excess load acting on the oxygen ions. Since $(d_h/d_o) > 1$, it follows that $\mu_x$ has a higher value than that corresponding to equilibrium conditions; the crystalline lattice reacts on this by expelling some of the oxygen ions, until the lattice is so poor in oxygen as to become crystallographically unstable; then it inverts into a $Fe_3O_4$-lattice,[1] or into a FeO-lattice, and eventually into metallic iron.

The task before us is, therefore, to evaluate numerically the increase with depth of the chemical potential of oxygen in layers, or earth shells, composed of $Fe_2O_3$, $Fe_3O_4$, and eventually FeO.

[1] The possible effect of the "reversed" specific-gravity relations of hematite and magnetite will not be discussed here.

The chemical potential at the surface has the following general relation to the vapor tension, $p$:

$$\mu = RT \cdot \ln p + \text{constant} ;$$

in all numerical calculations we shall use log $p$ as a measure for $\mu$. All the necessary constants are compiled in tables 1, 3, and 4.

The general form of the equation is:

$$\log p_x = \log p_0 + X \frac{M}{RT}\left(\frac{d_1}{d_2}\right) \log e \,. \quad (4)$$

At 500° C. we find the relation between depth, $X$, and the vapor tension, $p_x$, of oxygen in:

A layer of hematite: $\log p_x = X \cdot 6.76 \cdot 10^{-7} - 28$ ;

A layer of magnetite: $\log p_x = X \cdot 7.05 \cdot 10^{-7} - 30$ .

TABLE 3*

VAPOR TENSION EXPRESSED IN ATMOSPHERES†

| | 500° C. | 560° C. | 728° C. | 1150° C. |
|---|---|---|---|---|
| Oxygen in $Fe_2O_3$ ($p_0$) | $10^{-28}$ | $10^{-24}$ | $10^{-16}$ | $10^{-4.2}$ |
| Oxygen in $Fe_3O_4$ ($p'_0$) | $10^{-30}$ | $10^{-26.7}$ | $10^{-20}$ | $10^{-10}$ |
| Oxygen in FeO ($p''_0$) | Unstable | $10^{-26.7}$ | $10^{-20.7}$ | $10^{-12.6}$ |

* In this table the densities of oxygen have been computed from the ionic radii and the densities of the several iron oxides. In the following transformations the molecular volumes are shown:

$$3\,Fe_2O_3 \rightarrow 2\,Fe_3O_4 + O$$
$$\underbrace{3 \times 30.4}_{91} \rightarrow \underbrace{2 \times 44.5}_{89}$$

$$Fe_3O_4 \rightarrow 3\,FeO + O$$
$$44.5 \rightarrow 35.9$$

These equations indicate that, in the transformation hematite→magnetite and magnetite→wüstite, the removal of one oxygen ion corresponds to a decrease in volume of 2 units and 8.6 units, respectively. The two values are quite different, and only the last value compares with the data of table 4, from which the fictive volume of oxygen in magnetite is computed as

$$\frac{\text{Atom weight}}{d_o'} = \frac{16}{1.62} = 9.9 \,.$$

Thus, for the purpose of the present paper, which is mainly a discussion of the transformation, magnetite→wüstite, we can, without serious error, use the constants of table 4.

† Computed and extrapolated from Internat. Critical Tables, from Landoldt-Börnstein's Physicalische Tabellen, and from recent papers by: L. S. Darken and R. W. Gurry: The system iron-oxygen I and II: Jour. Am. Chem. Soc., vol. 67, p. 1398, 1945; vol. 68, p. 798, 1946.

TABLE 4

SPECIFIC GRAVITIES

| | | | | |
|---|---|---|---|---|
| Hematite ($d_h$) | 5.26 | Oxygen in hematite ($d_o$) | 1.71 | $d_h/d_o = 3.07$ |
| Magnetite ($d_m$) | 5.18 | Oxygen in magnetite ($d'_o$) | 1.62 | $d_m/d'_o = 3.20$ |
| Wüstite ($d_w$) | 6.00 | Oxygen in wustite ($d''_o$) | 1.65 | $d_w/d''_o = 3.64$ |

Molecular weight, $M$, of oxygen = 0.032 kg·mol⁻¹;
Gas constant $R$ = 82 kg. · cm · mol⁻¹ · degree⁻¹.

By using the above constants, the depth, $X$, is found in centimeters.

The condition of equilibrium between a surface layer of hematite and oxygen ions at a depth, $X$, is

$$\log p_x = X \cdot 2.20 \cdot 10^{-7} - 28\ .$$

These three curves are shown graphically in figure 2.

At 728° C. the corresponding equations are:

| | |
|---|---|
| In a layer of hematite: | $\log p_x = X \cdot 5.22 \cdot 10^{-7} - 16$ ; |
| In a layer of magnetite: | $\log p_x = X \cdot 5.45 \cdot 10^{-7} - 20$ ; |
| In a layer of wüstite: | $\log p_x = X \cdot 6.20 \cdot 10^{-7} - 20.7$ ; |
| The equilibrium condition: | $\log p_x = X \cdot 1.70 \cdot 10^{-7} - 16$ . |

At 1150° C. the corresponding equations are:

| | |
|---|---|
| In a layer of hematite: | $p_x = X \cdot 3.7 \cdot 10^{-7} - 4.2$ ; |
| In a layer of magnetite: | $p_x = X \cdot 3.84 \cdot 10^{-7} - 10$ ; |
| In a layer of wüstite: | $p_x = X \cdot 4.3 \cdot 10^{-7} - 12.6$ ; |
| The equilibrium condition: | $p_x = X \cdot 1.2 \cdot 10^{-7} - 4.2$ . |

The results of these computations demonstrate, theoretically, that the oxygen ions are in equilibrium with the hematite lattice only in the uppermost layer (disregarding again the zone of sedimentation and oxidation). As soon as the pressure becomes appreciable, magnetite will form.

The magnetite layer in a crust composed only of iron oxides will reach down to 41 km. at 500° C., to 107 km. at 728°, and to 220 km. at 1150°.

Now we shall leave the hypothetical iron oxide crust and discuss the conditions in an idealized lithosphere.

We assume that hematite is unstable at any depth below the zone of sedimentation and oxidation.

Magnetite is stable to greater depths. The specific gravity of the lithosphere at greater depths is approximately $d = 3.2$. Consequently, the ratio: $d/d_0 = 2.0$, and the calculated depth at which magnetite is in equilibrium with a top layer whose vapor tension is like that of hematite is:

| | |
|---|---|
| At 500° C., | 91 km., |
| At 728° C., | 235 km., |
| At 1150° C., | 483 km. |

At still greater depths, magnetite is unstable, and, if the temperature at these depths is as low as 500° C., then metallic iron will become stable at a depth exceeding 91 km. At temperatures above *ca.* 600° C. the magnetite zone will be underlain by a layer in which wüstite is stable, before one reaches the region of stable metallic iron.

### THE APPLICABILITY OF THE CALCULATIONS

The calculations of the various depth ranges are highly theoretical, for they are based on assumptions that correspond to

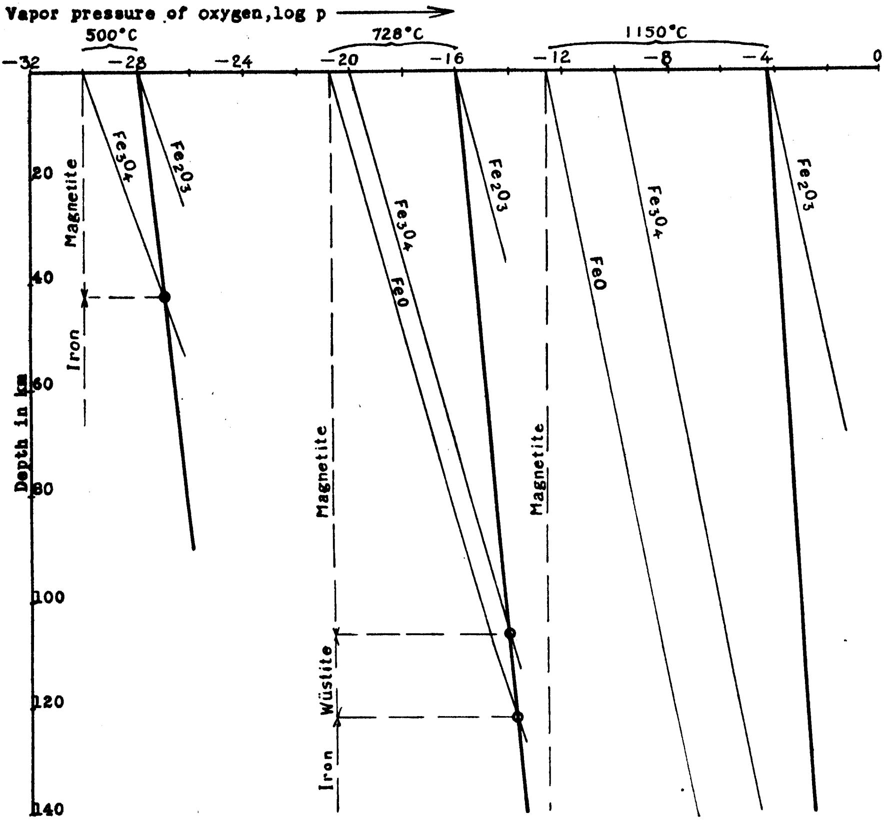



FIG. 2.—Graphical representation of the increase of oxygen pressure with depth, at three different temperatures, in hematite, magnetite, and wüstite. For each temperature is also given, in heavy lines, the curve representing the increase of oxygen pressure corresponding to equilibrium with a top layer of hematite. As seen from these curves, the oxygen tension in hematite increases with depth much faster than that which corresponds to the equilibrium tension. Therefore, hematite is theoretically unstable at any depth and inverts to (oxygen-rich) magnetite.

At 500° C. magnetite is stable to a depth of 41 km. At greater depth the oxygen tension in magnetite becomes too large, the crystalline lattice breaks up, and metallic iron becomes stable. (Wüstite, or FeO, is unstable at this temperature.)

At 728° C. the magnetite layer extends to a depth of 107 km. It is followed by a thin layer of wüstite in the range 107–122 km. At greater depth metallic iron is stable.

At 1150° C. the corresponding demarcation surfaces are at 220 km. and 305 km., respectively. These numerical values are correct only for a hypothetical crust composed of pure iron oxides.

idealized conditions never met with in nature. It is important, therefore, to discuss how far these calculations apply to natural conditions.

Obviously, many chemical elements bear no relation to the law of distribution implied in equation (1). Uranium, for instance, should be found in great depths only, but the opposite is the case. Uranium is a typical "lithophile" element, i.e., the chemical affinity relations of uranium bring it into granitic rocks rather than into the deeper strata, in spite of the gravity relations.

Now oxygen represents a different problem. Because of its great bulk in the lithosphere (= oxysphere) it can be regarded as a "solvent" capable of dissolving the various cations partly in the crystalline state but, at greater depths, in the liquid or glassy state.

Broadly, the "solute" (= all cations taken as an average) shows a tendency to distribution in accordance with equation (3). Thus, if one started with a homogeneous distribution, the tendency would be for the heavier cations to move downward, entailing a relative concentration of oxygen at the top. We have previously considered migration of oxygen upward; this migration is only relative. Owing to the great bulk of the oxygen ions, they probably furnish a rather stationary medium through which the smaller cations move. Through the migration of the cations a vertical oxygen gradient is established, and thermodynamic equilibrium is attained.

Instead of presenting the vertical gradient in terms of volume percentages of oxygen, as was done in figure 1, one might, with advantage, illustrate the same fact by comparing the total number of cations contained in equal rock volumes from various depth.

The data of the present paper make it clear that the volume relations of the ordinary rock types are almost wholly determined by the oxygen ions. If one wants to compare rocks of equal volumes, one can compare rocks containing the same amount of oxygen ions. As a standard comparison unit it is expedient to choose a volume comprising 160 oxygens, for in an ordinary rock nearly 100 cations are associated with 160 oxygens. We shall call a rock unit containing exactly 160 oxygen ions the "standard cell."

TABLE 5*

NUMBER OF CATIONS IN A CELL MADE UP OF 160 OXYGEN IONS

| | Number of Cations | Number of Oxygen Ions |
|---|---|---|
| Quartz | 80.0 | 160 |
| "Ichor" granite | 96.2 | 160 |
| Average igneous rocks | 99.3 | 160 |
| Average basalt | 102.4 | 160 |
| Olivine | 120.0 | 160 |

* The average rock analyses used in tables 1 and 2 are used here. The water content has been neglected, to eliminate the effect of the hydrogen ion.

A comparison of the standard cells of rocks of the various depth zones is afforded by table 5, which convincingly demonstrates the increase in cations with depth and thus, in different terms, illustrates the same thing as figure 1.

We conclude, therefore, that the distribution of oxygen in the lithosphere is roughly in accordance with equation (3) and that the mechanism of establishing the equilibrium is a migration of cations in a (glassy) solvent of relatively stationary oxygen ions.

Likewise, the distribution of the iron ions may be expected to be governed roughly by equation (3). Although many additional factors most certainly will bring about great local and regional deviations, it is a fact that iron is an element showing, at the same time, lithophile,

chalcophile, and siderophile tendencies and will therefore behave more like an "average" cation than many of the other elements.

### THE INORGANIC CYCLE OF OXYGEN

The discharge of oxygen from the deeper parts of the earth's crust is a geochemical process of great importance. Goldschmidt (1938) has regarded water, $CO_2$, and large quantities of Cl, F, B, S, and Se as products of degassing of the lithosphere. At one time oxygen also was formed through degassing.

Pertinent data have been published by Tammann (1924, p. 17), who demonstrated rather convincingly how great quantities of free oxygen must have developed immediately after the formation of a solid crust. At high temperatures and with a high partial pressure of steam in the atmosphere, the dissociation of steam was at that time great enough to provide for all the free oxygen now found in the atmosphere (Wildt, 1942, pp. 151–159).

Goldschmidt has criticized Tammann's quantitative analysis for the reason that through the geologic ages great amounts of oxygen became "fossil," i.e., deposited in oxidized sediments. Goldschmidt was impressed by the fact that the amount of chemically combined oxygen in the lithosphere was insufficient for a complete saturation of the highest valences of the cations (silicon and metals),[2] and he concluded that the primeval oxygen soon became "fossil" and that all oxygen in the present atmosphere was formed secondarily by photosynthetic reduction of carbon dioxide.

This can be elucidated by a simple computation. The mass of the atmosphere is approximately $5.1 \times 10^{15}$ tons; and the mass of free oxygen in the atmosphere is therefore only $1.15 \times 10^{15}$ tons. The mass of the lithosphere (16 km. depth) is $19{,}000 \times 10^{15}$ tons; in addition to 3.1 per cent $Fe_2O_3$, it contains 3.71 per cent, or $705 \times 10^{15}$ tons, of FeO.

Whether or not this large mass of ferrous iron is oxidizable, it is impossible to say. Part of it is present in magnetite and is not oxidizable, the largest part is combined in silicates, and the geological facts indicate that it is also not oxidizable.

However, if we make the assumption that all ferrous iron is available for oxidation according to the equation

$$2FeO + O = Fe_2O_3 ,$$

then the total free oxygen in the atmosphere would be sufficient to saturate $(1.15 \times 100)/78.5 = 1.5$ per cent of the lithosphere, corresponding to the very modest depth-range of 240 meters.

Assuming geologically stable conditions, a laterite weathering to a depth of 240 meters all over the globe is easily possible.

The attainment of this condition would deplete the air of all oxygen. Certainly, organic life is in a precarious situation. Perhaps this is now the condition of Mars with its red surface and deoxidized atmosphere. Fortunately, there are very few "fossil" or stationary situations in our good earth. Oxygen, in addition to its biochemical cycle, undergoes an equally important inorganic cycle.

Highly oxidized sediments and surface rocks are taken down to great depths by orogenic movements. Oxygen slowly re-

[2] Excerpt from Goldschmidt (1938, p. 27): "Dieses Defizit an Sauerstoff tritt zweifellos noch stärker in Erscheinung, wenn man die tieferen Teile des Erdballes mitberücksichtigt. Wir dürfen daher mit Bestimmtheit sagen, dass die Menge des Sauerstoffs im Gesamterdball nicht zur Absättigung der elektropositiven Elemente ausreicht, wobei die relativ kleine Gewichtsmenge des, sekundär entstandenen, freien Sauerstoffs der Atmosphäre vernachlässigt werden kann."

turns to the surface partly in rock minerals and partly combined in $H_2O$ or $CO_2$. Likewise, atomic oxygen would seem to regenerate by the dissociation of oxides at high temperature and pressure. Then new rocks at the surface are oxidized and again reduced at depth.

In each cycle a slight separation of the oxygen isotopes takes place; $O^{18}$ is preferentially retained in the mineral lattices of the deep-seated rocks; $O^{16}$ is delivered into the hydrosphere and atmosphere. Isotope determinations on suitably selected material would be of great geological interest.

ACKNOWLEDGMENT.—I wish to express my thanks to Mr. Terkel Rosenqvist, of the Institute for the Study of Metals of the Univeristy of Chicago, who has contributed by discussions of several problems, and by assembling the constants given in table 3.

## REFERENCES CITED

BARTH, TOM. F. W. (1948) The Birkeland granite, a case of petroblastesis: Soc. géol. Finlande Compte rendu, vol. 20 (Eskola vol.).

BROWN, H., and PATTERSON, C. (1947) The composition of the silicate phase of stony meteorites: Jour. Geology, vol. 55, pp. 405–411.

CLARKE, F. W., and WASHINGTON, H. S. (1924) The composition of the earth's crust: U.S. Geol. Survey Prof. Paper 127.

DALY, R. A. (1927) The geology of Saint Helena Island: Am. Acad. Arts Sci. Proc., vol. 62, no. 2.

GOLDSCHMIDT, V. M. (1928) Über die Raumerfüllung der Atome in Kristallen und über das Wesen der Lithosphäre: Neues Jahrb. f. Min., Beil. Band 57*A*.

——— (1938) Geochemische Verteilungsgesetze IX: Skr. Norske Vidensk. Akad. no. 4, 1937, Oslo.

RAMBERG, H. (1944*a*) Relation between external pressure and vapor tension of compounds, and some geological implications: Vidensk. Akad. Avh. Mat-Nat. Kl. 3, Oslo.

——— (1944*b*) The thermodynamics of the earth's crust. I: Norsk. Geol. Tidsskr., vol. 24, p. 98–111.

——— (1945) The thermodynamics of the earth's crust. II: *ibid.*, vol. 25, pp. 307–326.

——— (1946) Kjemisk likevekt i gravitasjonsfeltet, etc., with English summary: Chemical equilibrium in the gravitational field and some geological implications: Dansk. geol. Foren. Medd., vol. 11, pp. 13–29.

TAMMANN, G. (1924) Die Entstehung des freien Sauerstoffs der Luft: Zeitschr. physikal. Chemie, vol. 110, p. 17.

WILDT, R. (1942) The geochemistry of the atmosphere and the constitution of the terrestrial planets: Rev. Modern Physics, vol. 14, pp. 151–159.

# OXYGEN IN ROCKS: A BASIS FOR PETROGRAPHIC CALCULATIONS

TOM. F. W. BARTH
University of Chicago

ABSTRACT

The standard cell is defined as a rock unit containing 160 oxygens. The sum of the cations (silicon and metals) associated with this unit is very nearly 100 in all ordinary rock types. It is shown that, by using the standard cell as a basis, important petrogenetic relations can be surveyed and quantitatively defined more satisfactorily than by using any other method of petrographic calculation, including the norm method.

## THE STANDARD CELL OF A ROCK

In the preceding article on the distribution of oxygen in the lithosphere it was demonstrated that the distribution follows the theoretical laws of thermodynamic equilibrium demanding a regular decrease of oxygen with depth.

This fact can be presented in two ways: in terms of decreasing oxygen percentages or in terms of increasing numbers of cations associated with a fixed number of oxygens. In table 5 of the preceding paper a rock unit containing 160 oxygens was used as a reference standard; the table demonstrates that in near-surface rocks less than 100 cations are associated with this unit and that at greater depth more than 100 cations are present. The use of the so-called "standard" cell of 160 oxygens was proposed in an earlier paper on rock alteration (Barth, 1945, 1948). In the present paper I shall offer this method for general consideration to be used as a basis in petrographic calculations.

In most rocks oxygen makes up about 92 per cent by volume; all cations taken together (silicon and metals) make up but 8 per cent by volume. Consequently, the number of oxygen ions is of the utmost importance for the volume relations in rocks. Of secondary importance is the packing, and of still less importance are the number and kinds of the constituent cations.

In petrographic calculations it is often important to compare rocks of equal volume because most replacement processes in rocks and mineral deposits take place without appreciable change in volume, often with preservation of the most delicate structures. Likewise, rock weathering, alteration by hot springs, and similar processes are known to occur without much change in volume and with the original structures and textures surprisingly well preserved.

In view of what has been said in the preceding paper about the role of the oxygen ion in rocks and in mineral lattices, it is reasonable to suppose that the mechanism of such isovolumetric alterations is that of a migration and exchange of cations in a medium composed of relatively stationary oxygen ions. Only in this way can one explain the preservation of the delicate structural features; for removal of the large oxygen ions would break up the minerals and destroy the fragile and fine structure patterns. In harmony with this idea is, e.g., the mechanism of the weathering of biotite (bleaching, baueritization), which is effected by selective removal of the metal ions while oxygen and silicon remain in the residual lattice.

Thus, if we think that rock alterations take place without change in volume, it is equivalent to saying that the number of oxygen ions has remained constant. If

we want to compare isovolumetric rock units, then we should compare units containing the same number of oxygens. These considerations in their first approximation are restricted to rocks belonging to the same mineral facies.

The leptologic structure of most silicate rocks is such that on 160 oxygens there are always very nearly 100 cations. If we list separately the cations that are associated with 160 oxygen ions, the list will, therefore, very nearly, correspond to the atomic percentages of the rock-forming elements except oxygen, or, as we shall see presently, to the so-called "equivalent molecular percentages" as introduced by Niggli (1936, pp. 295–317) some years ago.

As a standard of comparison it is expedient, therefore, to choose a rock volume containing exactly 160 oxygen ions. A volume of this size I propose to call the "standard cell of a rock."

By referring the chemical composition of a rock to its standard cell we obtain a survey of important petrological relations not revealed by a simple inspection of the analytical data.

### OXYGEN AND THE ROCK-FORMING MINERALS

Before discussing the various rock types, it is important to investigate the relations of the rock-forming minerals. Pertinent data are collected in table 1. The chemical properties of the various minerals excellently illustrate the fact that oxygen and, of course, hydrogen show a strong tendency to accumulate near the surface and to become scarcer with depth.

In the minerals of the weathered sedimentary rocks few cations—about 80 or less—are associated with the standard cell, while the hydrogen content is very high; in the minerals of the deep-seated, metamorphic rocks the number of the cations in the cell attains 120, and no hydrogen is present. The cation concentration in all other mineral assemblages is between these two extremes.

This illustrates the geological cycle of oxygen in the lithosphere. We shall call it the "geological oxidation and reduction cycle." At the surface the minerals are *geologically oxidized*, which is mainly effected by the weathering agents, which remove the metal ions from the mineral lattices, often replacing them by hydrogen ions, thus entailing a relative abundance in oxygen. At great depths the minerals are *geologically reduced*, those rich in hydrogen and oxygen recrystallize, water is expelled, and a higher number of cations are introduced into the standard cell.

### NIGGLI'S METHOD

In an important work on petrographic calculations, Niggli (1936) rightly emphasizes the importance of a method by which one can rapidly survey the relations between the mineralogical and the chemical composition of a rock.

The calculation of the so-called "norm" is a tool which, particularly for igneous rocks, has proved its great value. But in metamorphic rocks, in which the interrelation of mineralogy and chemistry is of the utmost importance, the norm classification has failed. The reason is obvious: The normalization of the different metamorphic mineral facies encounters great difficulties in a norm based on weight percentages. The chemical relations of a rock in terms of weight percentages obscure the comprehensive view; the computations, moreover, become unnecessarily cumbersome.

*I propose, therefore, that calculation of the classical weight norm be altogether discontinued.* I regard it as obsolete and

## TABLE 1*

### Number of Cations in Rock-Forming Minerals
### (Referred to a Unit Cell of 160 Oxygens)

| Minerals of the Weathering and Zeolite Zone | | $\Sigma$ Cations | H |
|---|---|---|---|
| Kaolinite† | $Al_4Si_4O_{10}(OH)_8$ | 71 | 71 |
| Montmorillonite‡ | $Y_{2-3}Z_4O_{10}(OH)_2$ | 80 | 27 |
| Brucite | $Mg_3(OH)_6$ | 80 | 160 |
| Zeolite | $W_{5-12}Z_{40}O_{80} \cdot sH_2O$ | <80 | ∾120 |

| Epimetamorphic Facies | | $\Sigma$ Cations | H |
|---|---|---|---|
| Chlorite | $Y_3Z_2O_5(OH)_4$ | 88.9 | 71 |
| Talc | $Mg_3Si_4O_{10}(OH)_2$ | 93.3 | 26.7 |
| Muscovite | $KAl_3Si_3O_{10}(OH)_2$ | 93.3 | 26.7 |
| Zoisite | $Ca_2Al_3Si_3O_{12}(OH)_2$ | 98.5 | 12.3 |
| Actinolite | $X_2Y_5Z_8O_{22}(OH)_2$ | 100 | 13.3 |
| Albite | $NaAlSi_3O_8$ | 100 | 0 |

| Dynamometamorphic Facies | | $\Sigma$ Cations | H |
|---|---|---|---|
| Chloritoid | $FeAl_2SiO_5(OH)_2$ | 91.4 | 45.2 |
| Staurolite | $FeAl_4Si_2O_{10}(OH)_2$ | 93.3 | 26.7 |
| Muscovite | | 93.3 | 26.7 |
| Cyanite, Sillimanite | $Al_2SiO_5$ | 96. | 0 |
| Zoisite and epidote | | 98.5 | 12.3 |
| Alk.-hornbl. | $W_3Y_5Z_8O_{22}(OH)_2$ | 106.7 | 13.3 |
| Feldspar | $WZ_4O_8$ | 100 | 0 |
| Jadeite | $NaAlSi_2O_8$ | 106.7 | 0 |

| Mesometamorphic Facies | | $\Sigma$ Cations | H |
|---|---|---|---|
| Muscovite / Biotite | $KY_{3-4}Z_3O_{10}(OH)_2$ | 93.5 / 106.7 | 26.7 / 26.7 |
| Hornblende | $X_{2-3}Y_5Z_8O_{22}(OH)_2$ | 100.0 / 106.7 | 13.3 / 13.3 |
| Cordierite | $Mg_2Al_4Si_5O_{18}$ | 97.8 | 0 |
| Feldspars | $WZ_4O_8$ | 100 | 0 |
| Pyroxene | $XYZ_2O_8$ | 106.7 | 0 |
| Garnet | $X_3Y_2Z_4O_{12}$ | 120 | 0 |
| **Katametamorphic Facies** | | | |
| Spinel | $Y_3O_4$ | 120 | 0 |
| Olivine | $X_2ZO_4$ | 120 | 0 |

| Magmatic Facies | | $\Sigma$ Cations | H |
|---|---|---|---|
| Occasionally formed in magmas | | | |
| Feldspars | $WZ_4O_8$ | 100 | 0 |
| Pyroxenes | $XYZ_2O_6$ | 106.7 | 0 |
| Leucite | $KAlSi_2O_6$ | 106.7 | 0 |
| Melilite | $W_2YZ_2O_7$ | 114.3 | 0 |
| Nepheline | $NaAlSiO_4$ | 120 | 0 |
| Olivine | $X_2SiO_4$ | 120 | 0 |

* Quartz has the formula $Si_{40}O_{160}$ and may be present in all facies. In the table the letters W, X, Y, Z, designate ions of the following volume relations:

| | | | | | |
|---|---|---|---|---|---|
| Ion | $Si^{+4}$ | $Al^{+3}$ $Fe^{+3}$ | $Mg^{+2}$ $Fe^{+2}$ | $Na^{+}$ $Ca^{+2}$ | $K^{+}$ |
| Radius | 0.39 | 0.57 0.67 | 0.78 0.82 | 0.98 1.06 | 1.33 |
| | ←— Z —→ | | ←— X —→ | | |
| | | ←— Y —→ | | ←— W —→ | |

† Kaolinite, dickite, halloysite, nakrite.

‡ Montmorillonite, beidellite, nontronite, hectorite, saponite.

superseded partly by the Niggli "molecular norm," partly by the norm of the standard cell, as will be introduced in the present paper.

Niggli saw the great simplification in converting weight units into molecular units which he called "equivalent units" and which are defined as follows: Equivalent formula units are those which contain the same number of the constituent cations.[1]

Obviously, the equivalent units must be defined; for if one, e.g., writes "1 Nepheline" it may mean $NaAlSiO_4$ or $Na_2O \cdot Al_2O_3 \cdot 2SiO_2$, or $\frac{1}{3}(NaAlSiO_4)$, etc.

In $Na_2O \cdot Al_2O_3 \cdot 2SiO_2$ the sum of the cations would be $2Na + 2Al + 2Si = 6$. The "equivalent" unit of forsterite would be $4MgO \cdot 2SiO_2$ $(4Mg + 2Si = 6)$; and of enstatite, $3MgO \cdot 3SiO_2$ $(3Mg + 3Si = 6)$. It is simplest to reduce the several equivalent units to a sum of the cations $= 1$, then the sum of the several cations represents the number of equivalent units, thus (using the customary abbreviations, like Ne for Nepheline, etc.): $Na_2O \cdot Al_2O_3 \cdot 2SiO_2 = 6Ne$; $3MgO \cdot 3SiO_2 = 6En$; $MgSiO_3 = 2En$, etc.[2]

In this way the mineral equations become very simple, for instance,

$$Mg_2SiO_4 + SiO_2 = 2MgSiO_3 \ ,$$

$$3Fo + 1Q = 4En \ .$$

It is important to note that the sum of the coefficients of reaction on both sides of the equation is the same, for instance,

$$\underbrace{3Ne + 2Q}_{5} = \underbrace{5Ab}_{5}$$

Consequently, the sum of the molecular amounts remains constant. If a percentage distribution has been assumed, then any recalculation in terms of other combinations can be made without affecting the sum, which remains 100. This pertains to the norm calculation also. If a rock analysis is recalculated to a sum of 100 cations, then the molecular norm can be computed in practically no time. The molecular norm differs but slightly from the weight norm. This arises from the fact that the constituent oxides, $SiO_2$, $\frac{1}{2}Al_2O_3$, $CaO$, etc., have approximately the same weight. (Every petrographer knows, for instance, that the composition of a feldspar expressed in terms of Or, Ab, and An comes out with about the same figures, whether molecular percentage or weight percentage is used.)

Again I repeat my plea for using molecular units. An increasing number of petrographers in America and Europe alike, when publishing an analysis, calculate the corresponding norm values. It would be of great value if they would always apply molecular percentages. The fact that the molecular values are rather similar to the weight values makes the change easier.

### HOLMQUIST'S METHOD

A method that should not go into oblivion but can be used with advantage in modern studies is the molecular classification introduced by Holmquist in his great study of Swedish granites (1904–1905, pp. 78–269). At a time when the normative classification was introduced by the four leading American petrologists (Cross, Iddings, Pirsson, and Washington), Holmquist had made his own system. His principal point was that the chemical relations of a magma cannot be adequately discussed on the basis of weight percentages. In the treatment of chemical compounds and chemical reac-

[1] "Als übereinstimmende Formelgrössen können wir diejenigen bezeichnen, welche die gleiche Anzahl der wichtigen elektropositiven Elemente der Gruppe Si, Al, Fe, Mn, Mg, Ca, Na, K, Ti, Zr, Cr usw. enthalten" (Niggli, 1936, p. 296).

[2] Observe that in this scheme the symbol Ne has a quantitative meaning. It does not just mean the composition $NaAlSiO_4$ or $Na_2O \cdot Al_2O_3 \cdot 2SiO_2$, but it means a quantity of this composition corresponding to one cation.

tions, as well as in petrographic calculations, the weight units must be converted into molecular units.

More than thirty years before Niggli's publication, Holmquist recommended presentation of the rock analyses in terms of the constituent cations (metals and silicons) recalculated to 100 per cent. Like Niggli, he left oxygen out of consideration. The composition of a granite, for instance (from Halen, Sweden) may be written thus:

$$Si_{69.5}\,Al_{16.0}\,Fe_{1.4}\,Mg_{0.6}\,Ca_{1.4}\,Na_{5.2}\,K_{6.0}\ .$$

As the next step the cations can be rearranged as follows:

$$Si_{32.5} = 32.5\ ,$$

$$Si_{36.4}\,Al_{14.0}\,Ca_{1.4}\,Na_{5.2}\,K_{6.0}\,(+Al_{2.0}) = 63.0\ ,$$

$$Si_{0.6}\,Mg_{0.6} = 1.2\ .$$

$$Fe_{1.4}$$

This grouping is self-explanatory and give an idealized normative composition which compares well with the actual, or modal, composition, as seen in table 2.

TABLE 2

| Holmquist's Comp. | Mol. Per Cent | Modal Comp. | Weight Per Cent |
|---|---|---|---|
| Quartz | 32.5 | Quartz | 34.2 |
| Feldspar | 63.0 | Feldspar | 58.7 |
| Enstatite | 1.2 | Biotite | 3.7 |
| Fe, Al | 3.4 | Magnetite | 1.2 |

THE COMPOSITION OF THE STANDARD CELL

From the chemical analysis of a rock it is easy to calculate the elemental composition of the standard cell—it is simply to list the cations associated with 160 oxygen ions. The sum of the cations is near 100, and the figures, therefore, are nearly identical to the equivalent molecular percentages of Niggli and, furthermore, rather similar to the weight percentages as given by the chemists. This is important, because petrologists are accustomed to judge the composition of a rock by simply inspecting the weight percentages of the oxides. The numbers expressing the elemental composition of the standard cell are sufficiently close to those expressing the weight percentages of the oxides to allow us, generally speaking, to go by the old standards. This can be seen by inspection of table 3, which gives an example of the chemical composition of a hornblende minette expressed as (1) weight percentages of the oxides, (2) equivalent molecular percentages, and (3) number of cations associated with the standard cell. Table 4 gives, furthermore, the normative composition of the same rock in terms of (1) the usual

TABLE 3*

| | Weight Per Cent | Eq. Mol. Per Cent | No. of Cations in Standard Cell |
|---|---|---|---|
| $SiO_2$ | 52.68 | 51.6 | 49.4 |
| $TiO_2$ | 2.71 | 2.0 | 1.9 |
| $Al_2O_3$ | 14.02 | 16.2 | 15.6 |
| $Fe_2O_3$ | 5.04 | 3.7 | 3.5 |
| FeO† | 6.40 | 5.3 | 5.1 |
| MgO | 3.35 | 4.8 | 4.6 |
| CaO | 6.42 | 6.8 | 6.5 |
| $Na_2O$ | 2.43 | 4.6 | 4.4 |
| $K_2O$ | 3.30 | 4.1 | 3.9 |
| $P_2O_5$ | 1.30 | 1.0 | 1.0 |
| $H_2O$ | 1.89 | (6.2) | (5.9) |
| S | 0.60 | (1.0) | (1.0) |
| Sum | 100.14 | 100.0 | 95.9 |
| − for $S_2$ | 0.15 | ......... | ......... |

* The procedure of calculation is as follows: The figures giving the weight percentages are divided by the equivalent molecular weight of the corresponding oxide. Thus the figure for $SiO_2$ is divided by 60.06, the figure for MgO is divided by 40.32, and so on for all oxides containing one cation in the formula. For oxides containing *two* cations in the formula, we must divide by *one-half* the formula weight. Thus the figure for $Al_2O_3$ is divided by 50.97 (= ½101.97), the figure for $K_2O$ is divided by 47.1 (= ½94.2), etc. Thus we arrive at the cation proportions which, recalculated to 100 per cent, give the equivalent molecular percentages. Likewise, the figures for the proportions of the cations may be recalculated to give directly the number of cations in the standard cell. One just has to remember that there are ½, 1, 1½, or 2 oxygen ions associated with each individual cation and that the sum of the associated oxygens should be 160.

† Including 0.15 per cent MnO.

weight norm, (2) the Niggli molecular norm, and (3) mineral molecules in the standard cell. It is evident that the three different norms equally well accord to the petrologist a rapid survey of the potential mineral composition of the rock.

The "norm" of the standard cell as shown in the fourth column of table 4 is directly derived from the figures representing the cations in the cell (fourth col., table 3). Not only the norm but any mineral combination that one might like to express, normative or modal, can be exhibited by simply recombining the same figures. This advantage the standard-cell method shares with the Niggli method. In this respect the classical-norm method utterly fails.

TABLE 4

| | Weight Norm | Eq. Norm | Norm of Standard Cell |
|---|---|---|---|
| Q | 11.1 | 10.5 | 10.1 |
| Or | 19.6 | 20.5 | 19.6 |
| Ab | 21.1 | 23.0 | 22.1 |
| An | 18.2 | 18.7 | 17.9 |
| Wo | 2.8 | 2.8 | 2.7 |
| Fs | 3.1 | 2.7 | 2.6 |
| En | 8.4 | 9.6 | 9.2 |
| Ap | 2.8 | 2.6 | 2.5 |
| Il | 5.3 | 4.0 | 3.8 |
| Mt | 7.6 | 5.6 | 5.4 |
| Total | 100.0 | 100.0 | 95.9 |

How the cations of the standard cell can be recombined to give, quantitatively, the actual, rather complicated mineral composition of the rock, as qualitatively determined with the microscope, is demonstrated in table 5.

The chemical "formula" of the rock (analogously to the Holmquist method) can be presented as follows:

$$K_{3.9}\,Na_{4.4}\,Ca_{6.5}\,Mg_{4.6}\,Fe_{8.6}\,Al_{15.6}\,Ti_{1.9}\,Si_{49.4}\,P_{1.0}\,[O_{148.2}\,(OH)_{11.8}]_{160}\,.$$

However, the standard-cell formulas are, in one respect, fundamentally different from those proposed by Holmquist; they are based on a unit containing 160 oxygen atoms, whereas Holmquist disregarded the oxygen. Niggli likewise disregarded the oxygen, and therefore neither of these two methods can reflect the volume relations, which are of fundamental importance in geology.

### EXAMPLE I: DIABASE—HORNBLENDE MINETTE

The geological relations of the hornblende minette, the composition of which has been rendered in tables 3 and 4, present petrogenetic problems of general interest. The minette occurs 20 km. west of Kristiansand, on the southern coast of Norway (Barth, 1942). The mode of occurrence of the minette is that of a diabase dike apparently cutting pre-Cambrian migmatite of granitic composition; but a closer inspection shows that part of the dike is again cut by a pegmatitic facies of the migmatite. Consequently, the dike represents an old lamprophyre of pre-Cambrian age manifestly older than the last phases of the pre-Cambrian orogeny.

The mineral composition of the minette is obviously secondary, i.e., not magmatic but acquired through recrystallization. The analysis indicates that chemical alterations may have taken place at the same time. Briefly stated, the genetic problem is to investigate and define the metasomatic changes that took place during the metamorphism.

Without adducing further evidence it can be stated that there are reasons for

believing that the rock represents a diabase of the usual (basaltic) type, which later was changed into a hornblende minette.

Diabase dikes exhibit a remarkably homogeneous composition. Without any great error, therefore, we can assume that the pristine composition of the hornblende minette corresponds to the average analysis of all diabase dikes of the pre-Cambrian of southern Norway. Consequently, a comparison of the chemical composition of the original diabase with that of the minette should give us a clue to the changes. Such a comparison is difficult, however, if conservative petrographic methods are used. Some petrographers introduce entirely arbitrary assumptions, such as that iron or alumina is constant during the metasomatism. Now most petrologists agree that metasomatic changes take place volume for volume—and in some cases laborious calculations have been carried out to give educt and product the same volume. In the new method for calculation here proposed, the volume relations are automatically satisfied by referring everything to the standard cell. We compare the two standard cells directly, and the problem is solved without any trouble at all.

TABLE 5

| Si | Ti | Al | $Fe^{\cdot\cdot\cdot}$ | $Fe^{\cdot\cdot}$ | Mg | Ca | Na | K | P | (S) | ($H_2O$) | Modal Minerals |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 49.4 | 1.9 | 15.6 | 3.5 | 5.1 | 4.6 | 6.5 | 4.4 | 3.9 | 1.0 | (1.0) | (5.9) | 95.9 |
| 13.8 | | | | | | | | | | | | 13.8 Q |
| 7.5 | | 2.5 | | | | | | 2.5 | | | | 12.5 Or |
| 12.1 | | 4.1 | | | | | 4.1 | | | | | 20.3 Ab |
| 2.5 | | 2.4 | | | | 1.2 | | | | | | 6.1 An |
| 7.8 | 0.5 | 3.2 | 0.7 | 2.2 | 2.2 | 2.4 | 0.3 | | | | (1.2) | 19.3 Ho |
| 3.5 | 0.2 | 2.2 | 0.6 | 1.9 | 1.0 | | | 1.4 | | | (1.4) | 10.8 Bi |
| 1.0 | | 1.2 | | 0.2 | 1 4 | | | | | | (1.5) | 3.8 Chl |
| 1.2 | 1.2 | | | | | 1.3 | | | | | | 3.7 Ti |
| | | | | | | 1.6 | | | 1.0 | | | 2.6 Ap |
| | | | 2.2 | 0.8 | | | | | | (1.0) | (1.7) | 3.0 Ore |

DIABASE[3]

$$K_{1.4}\,Na_{4.5}\,Ca_{9.0}\,Mg_{5.9}\,Fe_{10.7}\,Al_{16.0}\,Ti_{1.9}\,Si_{48.5}\,P_{0.2}\,[O_{150.6}\,(OH)_{9.4}]_{160}\,.$$

MINETTE

$$K_{3.9}\,Na_{4.4}\,Ca_{6.5}\,Mg_{4.6}\,Fe_{8.6}\,Al_{15.6}\,Ti_{1.9}\,Si_{49.4}\,P_{1.0}\,[O_{148.2}\,(OH)_{11.8}]_{160}\,.$$

Since the two formulas represent the contents of the standard cells which have approximately the same volume, the difference between the rocks is found simply by subtracting the one from the other. Thus it is seen that the original diabase is metasomatically transformed into hornblende minette by addition and subtraction of the cations given in table 6.

Table 6 demonstrates that a remarkably small fraction of the rock substance (less than 1 per cent) need migrate in order to effect great metasomatic changes.

[3] Diabase dikes cutting pre-Cambrian rocks, southern Norway (average of 8). Recalculated from Barth-Correns-Eskola (1939, p. 72, table 24).

EXAMPLE 2: SOLFATARIC ALTERATION

In a chapter on "Petrography of Hawaii," Macdonald (Stearns and Macdonald, 1946, p. 192) describes solfataric alteration at Kilauea Caldera thus:

Gases escaping along faults at Kilauea Caldera have altered the adjoining rocks, both lavas and tuffs. Most prominent is a type of alteration caused by steam containing low concentrations of sulfuric, sulfurous, and carbonic acid. The resultant product is a rock composed largely of opal, with smaller amounts of kaolinite or related clay minerals, and relict magnetite and ilmenite. It has approximately the same volume as the unaltered rock, and original structures and textures are surprisingly well preserved.

TABLE 6

DIABASE PASSES INTO HORNBLENDE MINETTE BY

| Adding | Subtracting |
| --- | --- |
| 2.5 ions of K | 0.1 ions of Na |
| 0.9 ions of Si | 2.5 ions of Ca |
| 0.8 ions of P | 1.3 ions of Mg |
| 2.5 ions of H | 2.1 ions of Fe |
| | 0.4 ions of Al |
| Total: 6.7 cations representing 12.6 valences | 6.4 cations representing 12.6 valences |

These observations indicate that the number of oxygens must have remained constant during the alteration. Again we compare the standard cell of the parent-rock with that of the decomposed rock (recalculated from Payne and Mau, 1946, p. 351).

PARENT

$$K_{0.5}\ Na_{3.7}\ Ca_{10.6}\ Mg_{8.4}\ Fe_{7.0}\ Al_{17.0}\ P_{0.2}\ Ti_{1.8}\ Si_{48.8}\ O_{160.0}\ H_{8.4}\ .$$

DECOMPOSED

$$-\ -\ Ca_{0.1}\ -\ Fe_{0.2}\ Al_{0.1}\ -\ Ti_{6.4}\ Si_{55.9}\ O_{160.0}\ H_{69.6}\ .$$

Figure 1 illustrates graphically the changes indicated by the chemical analyses. Figure 1, *A*, is copied after Macdonald and Payne and Mau (1946, p. 351, table 3), plotted on the basis of no change in the silica content. This assumption is, of course, quite arbitrary; and figure 1, *B*, is believed to present a truer picture of what actually happened, plotted on the basis of no change in the oxygen content. It gives the following picture of the mechanism of the alteration: The effect of the acid hot-spring water is a selective removal of all metal ions and a simultaneous introduction of hydrogen, silicon, and titanium ions. Petrographic investigations in other hot-spring areas support this mode of interpretation.

EXAMPLE 3: THE STAVANGER AREA

The injection metamorphism of the Stavanger area, southern Norway, has become well known through Goldschmidt's work in 1920 and by the fact that several international excursions to the area have been organized. The main point of petrologic interest is the progressive metamorphism of the sediments by granitic and trondhjemitic intrusions.

The sediment by low-grade metamorphism is changed into a quartz-muscovite-chlorite phyllite; by increasing metamorphism the sediment is changed into albite porphyroblast schist and, further, into a granitelike rock. Goldschmidt gives the following discussions of the transition sediment → albite porphyroblast schist (pp. 113–114, translated from German):

In order to investigate the various possibilities of material transport we can calculate the composition of various rock mixtures, the soda content of which would correspond to that of the albite porphyroblast schist.

In this calculation the initial composition

is taken as the quartz-muscovite-chlorite phyllite.

The possibilities are as follows:

*a*) The material was introduced as trondhjemite; to 100 parts phyllite, 150 parts trondhjemite must be added.

*b*) The material was introduced as granite aplite; to 100 parts phyllite, 400 parts granite aplite must be added.

*c*) The material was introduced as albite; to 100 parts phyllite, 20 parts albite must be added.

*d*) The soda was introduced as albite; simultaneously silica and lime were introduced; water was removed. On 100 parts phyllite, 30 parts albite, 19.4 parts $SiO_2$, 3.9 parts Ca were added, 1.3 parts of water were subtracted.

*e*) The material was introduced and was *fixed* in form of oxides of $SiO_2$, $Na_2O$, CaO; water was removed; to 100 parts phyllite, 26 parts $SiO_2$, 3.1 parts CaO, 2.8 parts $Na_2O$ were added, 1.7 parts water were subtracted.

The results of these calculations prove that the introduction of soda did not occur as a simple addition of trondhjemite, granite, or albite; but that, according to the possibility *e*, from solutions or vapors lime and soda were selectively absorbed, with simultaneous precipitation of silica.

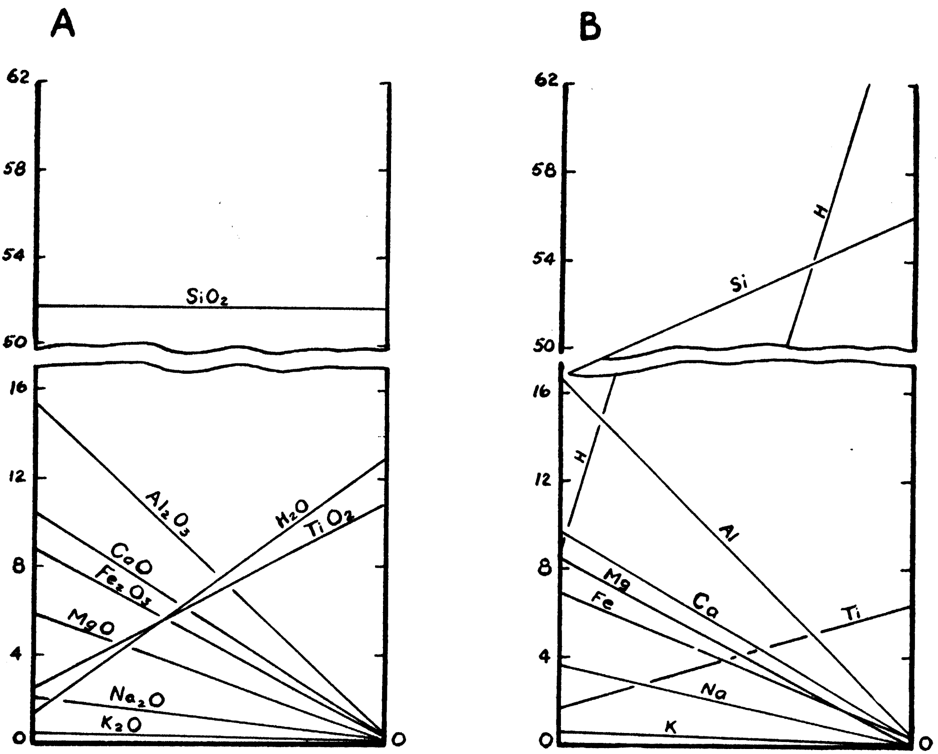



FIG. 1.—*A*, Alteration diagram according to earlier investigators; $SiO_2$ assumed to remain constant. *B*, Alteration diagram. Volumes are kept constant by keeping oxygen constant.

The foregoing paragraphs sum up the main conclusions and the main results of the paper by Goldschmidt, which at that time became well known and much discussed because it emphasized the importance of the functions exercised by the metasomatic solutions: the chemical composition of a metamorphic rock does

not generally correspond to a mixture of the sediment plus the injected "magma"; but from the magma certain oxides were selectively absorbed and fixed in the sediment, other oxides passed by without being fixed.

The volume relations were the chief reasons for Goldschmidt's rejecting the explanation of a simple mechanical mixing of the sediment and the intruded material. It is seen that the possibilities listed under *a* or *b* imply great quantities of foreign material to be introduced into the sediment. Goldschmidt's solution requires smaller quantities, although they are still appreciable. His solution can be represented by the following equation:

$$100 \text{ sediment} + 26.0\ SiO_2 + 3.1\ CaO + 2.8\ Na_2O - 1.7\ H_2O = \text{Albite-porphyroblast schist}.$$

Since the metamorphosed sediment shows no evidence of expansion, no swelling of the beds, etc., Goldschmidt's solution, which implies an increase of volume of more than 25 per cent, is not satisfactory. A metasomatism volume for volume is more in accordance with the field evidences. Therefore, we should compare the composition of the standard cells:

Sediment: $K_{5.1}\ Na_{2.1}\ Ca_{0.6}\ Mg_{2.4}\ Fe_{5.2}\ Al_{20.8}\ Ti_{0.6}\ Si_{51.0}\ C_{0.5}\ O_{160.0}\ H_{23.8}$.

Schist: $K_{3.9}\ Na_{5.3}\ Ca_{2.8}\ Mg_{1.9}\ Fe_{4.0}\ Al_{16.1}\ Ti_{0.4}\ Si_{57.1}\ C_{1.0}\ O_{160.0}\ H_{9.8}$.

Thus sediment passes into schist as shown in table 7.

TABLE 7

| Adding | Subtracting |
|---|---|
| 3.2 Na-ions | 1.2 K-ions |
| 2.2 Ca-ions | 0.5 Mg-ions |
| 6.1 Si-ions | 1.2 Fe-ions |
| 0.5 C-ions | 4.7 Al-ions |
| | 0.2 Ti-ions |
| Sum: 12.0 metal ions (34 valences) | 7.8 metal ions and 14.0 H-ions (34 valences) |

Table 7 shows that it is not sufficient to add soda, lime, and silica to the sediment to convert it to schist, as Goldschmidt claimed; there has to be a corresponding subtraction of other constituents.

Again the calculation of the chemical composition of the standard cells of the rocks gives us a much better understanding of the geological processes than does any other method of petrographic calculation.

## CONCLUSIONS

The standard cell of a rock is defined as a rock unit containing 160 oxygens. In most rocks oxygen makes up about 92 per cent by volume; the cations (silicon and metals) make up but 8 per cent by volume. Consequently, the number of oxygen ions is of great importance for the volume relations in rocks. If we want to compare isovolumetric rock units, then we should compare units containing the same number of oxygens.

In an average rock very nearly 100 cations are associated with the standard cell; consequently, a list of the cations will very nearly correspond to the atomic percentages of the rock-forming elements except oxygen.

The cations thus listed may be combined to form mineral molecules belonging to any one of the mineral facies (including, of course, the mineral molecules belonging to the conventional norm); any mineral combination that one might like to express, normative or modal, can

be exhibited by simply recombining the same figures.

It is a great simplification to use molecular units rather than weight units; therefore, the weight percentages should never be taken as basis for petrographic calculations. Equivalent molecular percentages are better to use; but, generally speaking, the best basis in petrographic calculations is the list of cations associated with the standard cell; this list represents the "chemical formula" of the rock, and on the basis of this list important petrologic relations can be worked out: (1) Degree of geologic oxidation. The cations in the standard cell, which—with the designation here adopted, correspond in number to the mineral molecules formed by combination of the cations—add up to a figure close to 100. In rocks of near-surface relations the number will usually be less than 100 (= more oxygen per cation: highly oxidized rocks), in deep-seated rocks the number will be more than 100 (less oxygen per cation: low degree of oxidation). (2) Most metamorphic and metasomatic processes in rocks take place without change in volume. In order to study such processes, for instance, in the study of compositional changes induced in a rock series through progressive metamorphism, the volume relations are automatically satisfied by referring the composition to the standard cell. By directly comparing the standard cells of the rocks under investigation, one obtains a rapid review of addition and subtraction of material caused by the petrogenetic processes. In this way the mechanism of granitization, the formation of "basic fronts," and similar phenomena can be studied more effectively than by any other method of petrographic calculation.

## REFERENCES CITED

BARTH, T. F. W. (1943) Lamprofyrer av to forskjellige aldre, etc.: Norsk. Geol. Tidsskr. vol. 23, p. 175.

——— (1945) The igneous rock complex of the Oslo region: Skr. Norske Vidensk. Akad., vol. 1, no. 8.

——— (1948) The nickeliferous Iveland-Evje amphibolite and its relations: Norges geol. undersökelse, vol. 168-*a*.

———, CORRENS, ESKOLA (1939) Die Entstehung der Gesteine, Berlin.

GOLDSCHMIDT, V. M. (1920) Die Injektionsmetamorphose im Stavangergebut: Vid. Akad. Skr. I, no. 10, Oslo, 1921.

HOLMQUIST, P. J. (1904–1905) Studien über die Granite von Schweden: Geol. Inst. Bull. 7, pp. 78–269.

NIGGLI, P. (1936) Über Molekularnormen zur Gesteinsberechnungen: Schweizer. min. pet. Mitt., vol. 16, pp. 295–317.

PAYNE, J. H., and MAU, K. T. (1946) A study of the chemical alteration of basalt: Jour. Geology, vol. 54, pp. 345–358.

STEARNS, H. T., and MACDONALD, G. A. (1946) Geology and ground-water resources of the island of Hawaii: U.S. Geol. Survey, Territory of Hawaii, Bull. 9.

# ISOTOPE RATIOS: A CLUE TO THE AGE OF CERTAIN MARINE SEDIMENTS

FRANS E. WICKMAN
University of Chicago

ABSTRACT

If an element $A$ has a radiogenic isotope, $A_1$, and a nonradiogenic one, $A_2$, the ratio $A_1/A_2$ is an index of the age of marine chemical sediments, if the content of the isotope $B^*$ producing $A_1$ can be neglected. It is shown that the method can be used for strontium (and perhaps for $Pb^{208}$) in limestones and anhydrites.

## INTRODUCTION

Age determinations have hitherto been performed on igneous rocks only; and this is unfortunate, as the relative geological time-scale is based on the sedimentary rocks, with their content of invertebrate fossils. Hence it may be of interest to attempt to develop methods suitable for sedimentary rocks.

## PRINCIPLES OF THE NEW METHOD

If a radioactive isotope generates an isotope of another element, the abundance of the isotopes of the daughter-element will vary with time, the abundance of the isotopes, and the disintegration constant of the radioactive isotope. If we consider the earth's crust, the isotopic composition is a special, evaluable function of time; but this statement is not true for a small part of it. The function will depend on the composition of the sample, its geological history, etc.

Let us now look at the problem from a different point of view. We can ask if there is any geological milieu about which we can be certain that the isotopic composition of an element is a one-valued function of time. I think that the sea is such a milieu. The isotopic composition of sea water is a one-valued function of time and probably represents, to a high degree of approximation, an average sample of the earth's crust.

If there is a difference between the earth's crust and the sea, the maximum would exist under the very special topographical conditions during which large fractions of the land areas do not drain to the sea. But even in this case I think the approximation to the earth's crust must be very close.

If a chemical sediment is precipitated in the sea, the isotopic constitution of the radiogenic element will also be a one-valued function of its age, if the content of the radioactive parent can be neglected. This means that the isotope ratios, at least in principle, can be used as age indicators and that the index multiplied by a constant gives the age of the sediment in millions of years.

The most suitable sediments seem to be limestones, anhydrites, and gypsum. In these cases strontium and lead are the most promising elements. The other disintegration products, such as helium, argon, calcium, neodymium, hafnium, and rhenium, are of no significance in this connection. The special problems of strontium and lead will be discussed separately.

## STRONTIUM IN LIMESTONES AND ANDHYDRITES

Strontium should be particularly favorable, as it is an important constituent of sea water, 13,000 $\gamma$/liter; and, on the other hand, the content of rubidium, which is the parent, is low, namely, 200

$\gamma$/liter (Sverdrup, Johnson, and Fleming, 1942). As W. Noll has shown (1934), strontium can replace calcium in calcium minerals such as calcite, dolomite, aragonite, gypsum, and anhydrite; and rubidium is, on the other hand, known not to enter these minerals. Strontium occurs most frequently in aragonite and anhydrite (table 1).

TABLE 1

VARIATION IN CONTENT OF STRONTIUM IN CERTAIN SEDIMENTARY MINERALS AND ROCKS ACCORDING TO W. NOLL (1934)

| Mineral, Rock | Per Cent SrO |
|---|---|
| Gypsum | 0.003 –(0.13) |
| Anhydrite | 0.17 – 0.69 |
| Calcite | 0.0005– 0.14 |
| Aragonite | Up to 4.69 |

TABLE 2

STRONTIUM CONTENT OF RECENT MARINE ORGANISMS, AFTER NOLL (1934)

| | Per Cent SrO |
|---|---|
| Corals: | |
| *Porites clavaria*, Florida | 0.5 |
| *Corallium rubrum*, Sicily | 0.2 |
| *Millepora alcicornis*, Florida | 0.5 |
| Cephalopod: | |
| *Nautilus pompilius*, Pacific Ocean | 0.4 |
| Brachiopod: | |
| *Terebratula vitrea*, Naples, Italy | 0.05 |
| Lamellibranchiata: | |
| *Pinna squamosa*, Naples, Italy | 0.04 |
| Gastropod: | |
| *Bulla ampulla*, Atlantic Ocean | 0.19 |
| Scaphopod: | |
| *Dentalium*, North Sea | 0.2 |
| Lithothamnium: | |
| *Lithothamnium polymorphum*, Adriatic Sea | 0.3 |

The high strontium content of gypsum refers to rock-forming gypsum, which is a product of metamorphism of anhydrite. Noll found that the content of strontium in pure gypsum crystals was very low. Thus it is probable that strontium occurs as celestine in rock-forming gypsum. In recent marine organisms the content of strontium can rise relatively high, as table 2 shows.

From these figures it must be concluded that the shells themselves can be used for age determinations. In this case the question arises: Are the strontium isotopes selectively concentrated by organisms? I think it is very improbable that a biological separation in this case would be of any importance at all; but in any case it is simple to investigate shells of recent marine organisms. Celestine is regularly found associated with fossils which have had shells of aragonite. Noll cites several examples from the literature (1934, pp. 574–575). Thus we must discuss one of the serious problems confronting this method. Are the isotope ratios changing during the diagenesis by the mixing of strontium atoms of different origin? This question can be answered only by an investigation of a thick lime deposit; but, nevertheless, it is of interest to give the opinion of Noll, especially in the case of crystals of celestine associated with fossils. He says (1934, pp. 575–576):

> Zu demselben Resultat: Verknüpfung der Coelestinanreicherungen mit Aragonittieren kam Liebetrau bei Untersuchungen in Thüringer Muschelkalk. Coelestin vertritt nach ihm ganz besonders gern die Substanz von Gastropodenschalen, nur untergeordnet von Lamellibranchiaten und nicht von Brachiopodenschalen. Es sind aber eben Gastropoden und Lamellibranchiatenschalen, die aus Aragonit bestehen. Liebetrau sieht die Erklärung für die Verknüpfung des Coelestins mit den Aragonithartteilen in der grösseren Löslichkeits des Aragonites gegenüber Calcit und damit der grösseren Umsatzfähigkeit des Aragonites mit zirkulierenden Lösungen. Damit dürfte aber noch nicht alles erklärt sein. Durch die Auflösung des Aragonites kann zwar ein Hohlraum entstehen, in dem Coelestinkristalle sich frei ausbilden können, *dass* sie

sich aber abscheiden, muss einen anderen Grund haben, der nach dem Massenwirkungsgesetz entweder in Erhöhung der $Sr^{2+}$ oder des $SO_4^{-2}$-Ionenkonzentration gesucht werden muss. Die Konzentration des $Sr^{2+}$ kann offenbar im Bereich der Auflösung der Aragonitschale erhöht werden, da die Aragonitteile Strontium angereichert enthalten. Der Angelpunkt für die Erklärung der Verknüpfung von Coelestin mit Aragonitschalen wäre also darin zu sehen, dass der hohe Strontiumgehalt des Aragonites durch dessen leichte Löslichkeit leicht freigemacht und im entstandenen Lösungshohlraum sogleich wieder abgeschieden werden kann.

Hence Noll concludes that there is very little mixing of material. This is, of course, especially the case when thin beds of limestone are separated by shales or other dense materials. Whether the method can be used will depend on the magnitude of the variations of the isotope ratios. In this respect we must know the content of $Rb^{87}$, $Sr^{87}$, and $Sr^{86}$ (or another nonradiogenic isotope of strontium) in the earth's crust, and the disintegration constant of $Rb^{87}$.

According to V. M. Goldschmidt (1937*b*), the content of rubidium is 310 gm/ton, and 28 per cent of this amount is radioactive rubidium 87. Thus the content of $Rb^{87}$ in the earth's crust is 86 gm/ton. The disintegration constant is, according to a new determination by S. Eklund (1946), $3.8 \pm 0.7 \times 10^{-19}$ $sec^{-1}$.

The strontium content in the earth's crust is much more difficult to estimate. According to Noll (1934), the content of SrO would be 0.05 per cent. His estimate was not made according to the common method of estimating the abundances of elements. Goldschmidt (1937*a*) uses the value 150 gm. Sr per ton. This value is probably low; a plausible value is 200–300 gm. Sr per ton. Using the higher value for our calculation, 300 gm. Sr per ton, and an isotope abundance of $Sr^{87}$ of 6.6 per cent, the present content of $Sr^{87}$ in the earth's crust would be 20 gm/ton. The present ratio $Sr^{87}/Sr^{86}$ is arbitrarily fixed as 0.6600. In figure 1 the strontium ratio is plotted on a geological time-scale. The lengths of respective periods are given according to A. Holmes (1947).

From this figure it is quite clear that the measurement of the isotope ratio $Sr^{87}/Sr^{86}$ must be made as accurate as possible. In order to get the opinion of an expert, I have discussed the problem with Dr. Alfred C. O. Nier, of the University of Minnesota, and he writes, in a letter to the present writer (published with the kind permission of Dr. Nier):

> There is little doubt in my mind that variations of 0.1% in the ratio can be detected if sufficient care is taken, and in time the precision may be extended to permit measurements to within 0.01%. Certainly there is no fundamental reason why a mass spectrometer could not be perfected to give such accuracies. Naturally, for some time to come it will be difficult, if not impossible, to achieve an accuracy of 0.01% on a routine basis, but if enough effort is expended it should be possible to obtain this accuracy as a regular thing.

This means that in the future the ratio can be given to four places, and the error will be about 0.00006. In the present case this is equivalent to the statement that the ages can be given with an error of about 1,000,000 years, regardless of their position on the time-scale. This error refers to the relative error and the relative time-scale, and probably it will be very difficult to determine the relation between the absolute age and the strontium ratio with the same accuracy.

### LEAD IN LIMESTONES AND ANHYDRITES

If the strontium content of sea water is high, the lead content is very low. According to M. Boury (1938), the lead content is about 4 γ/kg sea water. The uranium content, on the other hand, is 1.5 γ/kg, and the thorium content much

| Time | | $\frac{Sr^{87}}{Sr^{86}} \times 10^4$ | $\frac{Pb^{208}}{Pb^{204}} \times 10^3$ |
|---|---|---|---|
| 0 | | 6600 | 40 000 |
| 12 | PLIOCENE | 6595 | 39 992 |
| 26 | MIOCENE | 6588 | 39 982 |
| 38 | OLIGOCENE | 6583 | 39 973 |
| 58 | EOCENE | 6574 | 39 959 |
| 127 | CRETACEOUS | 6543 | 39 911 |
| 152 | JURASSIC | 6532 | 39 893 |
| 182 | TRIASSIC | 6519 | 39 873 |
| 203 | PERMIAN | 6510 | 39 858 |
| 255 | CARBONI-FEROUS | 6486 | 39 821 |
| 313 | DEVONIAN | 6461 | 39 780 |
| 350 | SILURIAN | 6444 | 39 755 |
| 430 | ORDOVICIAN | 6409 | 39 699 |
| 510 | CAMBRIAN | 6373 | 39 643 |

| Time | | $\frac{Sr^{87}}{Sr^{86}} \times 10^4$ | $\frac{Pb^{208}}{Pb^{204}} \times 10^3$ |
|---|---|---|---|
| 0 | | 6600 | 40 000 |
| 58 | | 6574 | 39 959 |
| 182 | | 6519 | 39 873 |
| 510 | | 6373 | 39 643 |
| 1000 | | 6155 | 39 300 |
| 1500 | | 5932 | 38 950 |
| 2000 | PRE-CAMBRIAN | 5710 | 38 600 |
| 2500 | | 5487 | 38 250 |
| 3000 | | 5265 | 37 900 |
| 3350 | | 5100 | 37 655 |

FIG. 1.—The variation of the ratios $Sr^{87}/Sr^{86}$ and $Pb^{208}/Pb^{204}$ with time

less than 0.5 $\gamma$/kg. These two measurements were made by H. Pettersson and his co-workers (1939).

Measurements of the lead content of marine sediments are very rare. According to Siebenthal (1915, pp. 78–79), in Cambrian and Ordovician magnesian limestones in Missouri the lead content varies between less than 4 to 16 gm/ton, with a mean of 10 gm/ton. In Mississippian limestones in Missouri, on the other hand, the content was about 20 gm/ton and in different limestones and dolomites from Iowa about 32 gm/ton. Forchhammer[1] found in 1865 the lead content in corals to be as follows:

| | Gm/Ton |
|---|---|
| *Pocillopora alcicornis* | 2.7 |
| *Heteropora abrotanoides* | 20 |

Of course, these figures are not of any great value, but they fit into what can be expected. As the ionic radii of $Sr^{2+}$ and $Pb^{2+}$ are very similar, it is probable that the lead and strontium content in sea water and marine chemical sediments are proportional. If this is the case, the lead content of limestones must be expected to be up to 10 gm/ton and in anhydrites up to 1 gm/ton.

Hence this method can be dangerous if the content of radioactive elements is relatively high. For electrostatical reasons it is not probable that uranium and thorium can occur in the crystal structure of calcite, aragonite, anhydrite, and gypsum. This is also in agreement with experience; Beer and Goodman write, for instance (1944, p. 1251):

> A pure, crystalline limestone or a pure, white quartz sand exhibits practically no measurable radioactivity. Those limestones which show appreciable activity also contain proportionate amounts of shale, organic matter, or other material which has once occupied the colloidal phase.

[1] Cited in Boury (1938).

Without any opinion on the correctness of their theory, it is nevertheless clear that the radioactivity of pure marine limestones is low. Probably it is advisable to use only the ratio $Pb^{208}/Pb^{204}$, that is, the ratio referring to the thorium series, as the thorium content of sea water is extremely low and the presence of organic matter can be an indication of an enrichment of uranium.

We assume the thorium content of the earth's crust to be 4 gm/ton and the dis-

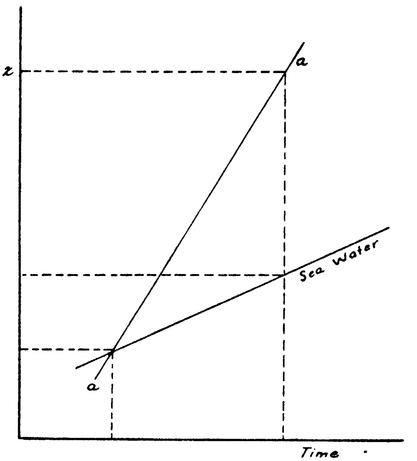



FIG. 2.—The variation of the ratio $Pb^{206}/Pb^{204} = x:1$ with time in limestone and sea water.

integration constant to be $4.99 \times 10^{-11}$ year$^{-1}$. The lead content is 20 gm/ton, and the percentage of $Pb^{208}$ is about 52. Thus the amount $Pb^{208}$ can be estimated to be 11.5 gm/ton. These values give the ratios $Pb^{208}/Pb^{204}$ plotted in figure 1.

Assuming the same accuracy as in the case of strontium, the error can be estimated to be about 7,000,000 years.

### LIMESTONES WITH HIGH CONTENT OF URANIUM

If the limestones contain uranium, then the present method cannot be used. On the other hand, the usual lead method

cannot be used, as the original composition of the common lead is not known. But, with some modifications, it is possible to use a method proposed by the present writer several years ago (1943). The principle is shown in figure 2. If we determine the isotope ratio of the limestone, we get the ratio 1:$x$. From the content of uranium and lead in the limestone, we can recalculate the variation of the isotope ratio with time (curve $a$). On the other hand, we know that the isotopic composition of the sample at the time of deposition was the same as the isotope composition of the earth's crust. This last curve is also known, and the point of intersection gives the age. If this method is to be accurate, the ratio of uranium to lead must be either high or low in relation to the ratios in the earth's crust, as can easily be seen from the figure.

## CONCLUSIONS

1. It is shown that the ratio between a radiogenic and a nonradiogenic isotope of an element can be used as an index of the age of marine chemical sediments, if the content of the mother-element can be neglected.

2. The method is especially suitable for strontium in limestones and anhydrites. It will be possible to determine the ratio with an error of 0.01 per cent. In order to get accurate absolute age determinations, the contents of rubidium and strontium in the earth's crust and the disintegration constant of $Rb^{87}$ must be determined with greater accuracy. The error of an age determination can be estimated to be 1,000,000 years.

3. It might be possible to use lead in some cases to check the results obtained with strontium ratios. The accuracy to be expected is less than in the case of strontium, perhaps 7,000,000 years.

4. A variation of the method proposed by the present writer (1943) is shown to be of some value in the determination of the age of marine limestones either rich or poor in uranium.

## REFERENCES CITED

Beers, R. F., and Goodman, Clark (1944) Distribution of radioactivity in ancient sediments: Geol. Soc. America Bull. 55, p. 1229.

Boury, M. (1938) Le Plomb dans le milieu marin: Revue des travaux de l'office des pêches maritimes, vol. 11, p. 157.

Eklund, S. (1946) Studies in nuclear physics; radioactivity of $Rb^{87}$: Arkiv Mat. Astron. Fys., vol. 33, no. A14.

Föyn, Ernst; Karlik, Berta; Pettersson, Hans; and Rona, Elisabeth (1939) The radioactivity of sea water: Meddelelser Oceanographic Inst. Göteborg, no. 2.

Goldschmidt, V. M. (1937$a$) Geochemische Verteilungsgesetze der Elemente: Skr. Norske Vidensk.-Akad. Oslo, vol. 1, Math.-natur. vid. Kl. no. 4.

——— (1937$b$) The principles of distribution of chemical elements in minerals and rocks: Chem. Soc. London Jour., p. 655.

Holmes, A. (1947) The construction of a geological time-scale: Geol. Soc. Glasgow Trans., vol. 21, p. 117.

Liebetrau, E. (1899) Beiträge zur Kenntnis des unteren Muschelkalkes bei Jena: Zeitschr. Deutsch. geol. Gesell., vol. 41, p. 721.

Noll, W. (1934) Geochemie des Strontiums; mit Bemerkungen zur Geochemie des Bariums: Chemie der Erde, vol. 8, p. 507.

Siebenthal, G. E. (1915) Origin of the zinc and lead deposits of the Joplin region, Missouri, Kansas, and Oklahoma: U.S. Geol. Survey Bull. 606.

Sverdrup, H. V.; Johnson, M. W.; and Fleming, R. H. (1942) The oceans, their physics, chemistry, and general biology, New York, Prentice-Hall Book Co.

Wickman, F. E. (1943) Can the "lead method" be used on igneous rocks? Arkiv Kemi, Mineralogi och Geologi, vol. 16, no. A23.

# THE PRESERVATION OF ANTARCTIC ICE SPECIMENS[1]

ARTHUR DAVID HOWARD
Stanford University

ABSTRACT

Small ice specimens, including thin sections and delicate ice crystals, were brought back intact from the Antarctic. The samples were kept in airtight containers and were stored in a cold-storage room aboard ship during the journey home. Thin sections were housed in slide boxes in which the sections rested horizontally. Immersion of the sections in kerosene to prevent evaporation was not entirely satisfactory because of the tendency of the kerosene to penetrate the section. However, such penetration can be prevented. The use of cover glasses with a seal of vaseline around the edges was found by European glaciologists to prevent evaporation. Specimens other than thin sections were packed in snow. The snow prevented jostling and served as an insulating medium.

The feasibility of transporting small specimens of ice over distances of thousands of miles in journeys involving months should encourage more comprehensive sampling of ice terrains. Furthermore, if detailed ice studies are impracticable in the field because of insufficient time, lack of necessary equipment, or lack of specialized personnel, specimens can be returned home for study.

## INTRODUCTION

The difficulties involved in the preservation of ice specimens are partly responsible for the slow progress in the study of the physical characteristics of this interesting rock. These difficulties are particularly serious in places like Antarctica, where field parties are frequently away from base for weeks or months at a time and the problems of space and weight vie in importance with that of preservation of specimens. Yet, until methods are devised whereby small ice samples can be transported over large distances with nearly the ease with which other rocks are handled, the variety of specimens needed for comprehensive ice studies will be lacking. Only when the problems of transportation have been solved will specialists in well-equipped laboratories at home have access to samples from diverse and distant environments. As it is, expedition geologists are often compelled by circumstances to embark on ice studies in spite of the fact that their principal interests lie in other directions.

As a member of the recent Navy Antarctic Expedition, 1946–1947, under the technical command of Rear Admiral Richard E. Byrd, the writer was able to conduct a few simple experiments bearing on this problem. The experiments were designed to indicate the minimum size of ice specimen that needs to be collected in order to insure safe transport back to the United States. Unfortunately, there were no long-range field trips during the present expedition, the entire study being conducted at Little America and in the ship's refrigerator en route home. The tossing of the ship during the stormy return passage, however, subjected the specimens to gyrations probably fully as bad as those that are encountered on sledges or tractors in the field.

The brief period of preparation available to the writer before the departure of the expedition, as well as the need for consideration of other problems, allowed little time for contemplation of methods and acquisition of equipment for this study. Methods were devised en route, and makeshift equipment was salvaged or built from odds and ends aboard ship.

[1] Published by permission of the director, United States Geological Survey.

Hence no pretense at thoroughness is made. Recommendations are presented as guides for further study.

Most of the specimens herein described are now in storage in a refrigerating plant in Washington, D.C. They will be examined periodically for further wasting.

## EQUIPMENT

Cylindrical tin cans of assorted sizes, with large circular lids, were selected for airtight storage (pl. 1, *A*). All paint was removed from the cans to permit maximum reflection of incident radiation. The large lids permitted the storage of specimens nearly as wide as the diameter of the cans.

A framework was built into one of the larger cans to house a specially built case for thin sections (pl. 2, *A*). This case was capable of holding, in horizontal position, ten 2×3-inch glass slides at intervals of ½ inch. Two small blocks of ice cemented to glass slides were also stored in this container. Plate 2, *A*, also shows a second slide box, not airtight, which provided a basis for comparison of results with the airtight container. All containers were stored in a large wooden case, and the interstices were packed with snow. This wooden box was sheltered under an outdoor work bench at Little America (pl. 1, *A*) and later was stored in the ship's refrigerator.

During the stay on the ice, all thin sections were stored in the metal, airtight container. The container was filled with kerosene, which has proved to be an effective deterrent to evaporation (Demorest, 1940, p. 27). When the container was transferred to the ship, however, it became necessary to discard the kerosene for fear of imparting an odor to the foodstuffs in the refrigerated room.

## TEMPERATURE CONDITIONS OF STORAGE

The maximum temperature recorded on the ice during our stay was 28° F. Thus, at no time were the specimens subject to melting temperatures. Other expeditions, however, have recorded temperatures above the melting-point at Little America. Although these were rare occasions of short duration, collection of specimens of somewhat larger size than would otherwise be collected might be advisable. The melting could be kept to a minimum or eliminated, however, by packing the containers in snow, which serves as a good insulator.

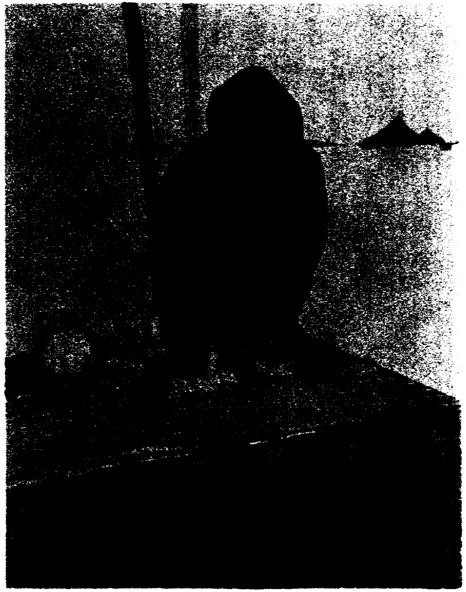
On shipboard the specimens were kept in a refrigerated storage room at a temperature of about 16° F. The ice studies which began at the base camp were continued in this room, something that could not have been done if the specimens had been stored in a small deep-freeze unit and a refrigerated room had been unavailable.

## THE SAMPLES

Samples used in the study included specimens of the névé of the Ross Shelf Ice, ice from crevasses in the pressure folds of the Bay of Whales, and delicate sublimation crystals from the interior of the buildings of Little America II and III.

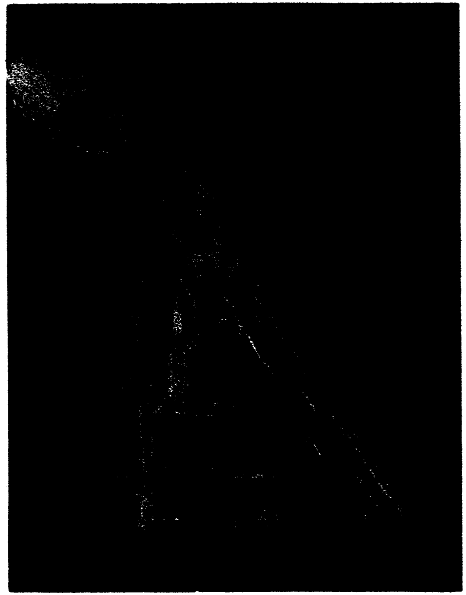
The density of the névé samples ranged from a little over 0.4 to about 0.54. Pore spaces were large, numerous, and frequently connected, allowing relatively easy penetration of air to a depth of probably 1–2 mm. One, and possibly both, of the crevasse specimens had a density of 0.87. Pore spaces were smaller, fewer, and more isolated than in the névé; hence these specimens exposed less surface to evaporation. No density determinations of the ice crystals were made.

PLATE 1



A

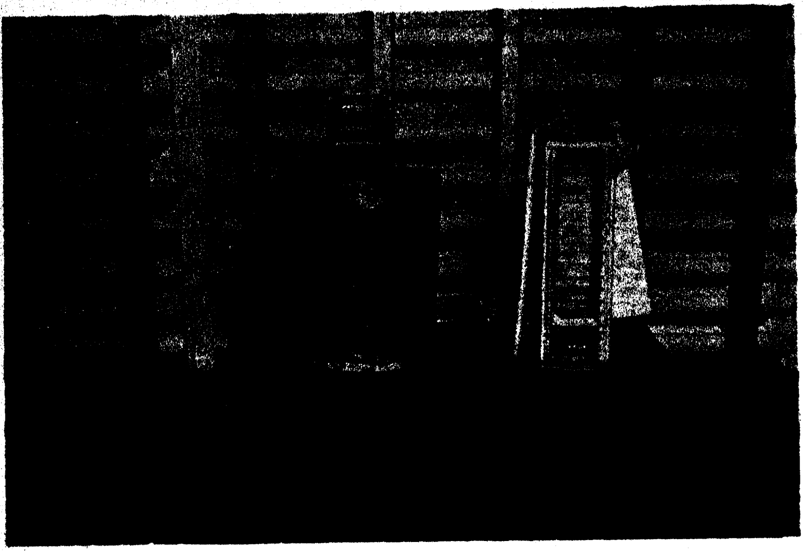
*A*, Work bench at Little America IV. Note the type of can used for airtight storage of specimens. The snow packing was used for all specimens except those mounted on glass slides. The double-bladed hacksaw, with blades 1½ inches apart, was designed by the writer to speed the preparation of thin sections 1½ inches square.



B

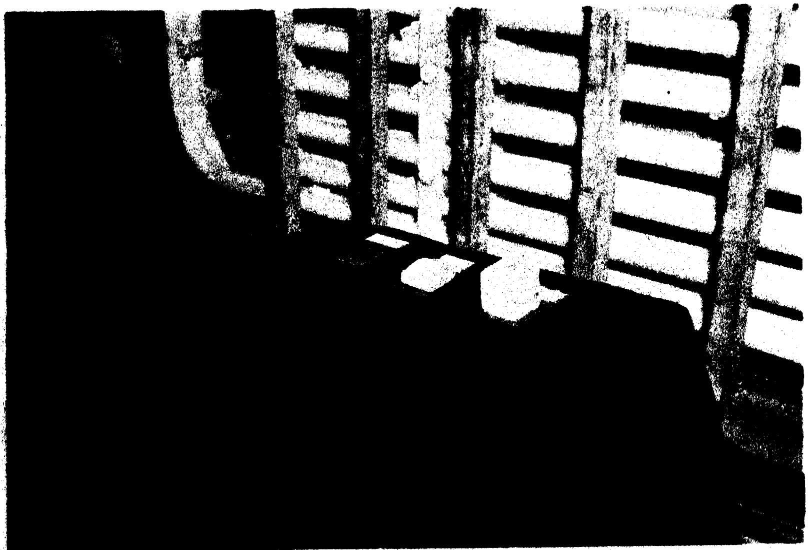
*B*, One of four sublimation crystals brought back from Little America. Length along diagonal, nearly 5 inches. This specimen, which showed no discernible evidence of wasting when examined in Washington, unfortunately broke while being prepared for rephotographing. The several fragments, as well as three other crystals, are in storage in Washington.

PLATE 2



A

*A*, The metal can served for airtight storage of thin sections. A housing within the can holds the slide-box snugly and prevents it from moving about. When in place, the slide-box rests entirely within the can. The box on the right is not airtight. Preservation results were compared with those obtained with the airtight can.



B

*B*, Thin section and two blocks fashioned from pressure ice from the Bay of Whales in late January, 1947. Photographed on arrival in Washington, April 14, 1947. The specimens suffered practically no deterioration in transport.

Twelve thin sections, $1\frac{1}{2}$ inches square and 0.5–1.0 mm. thick, which had been prepared for petrographic study, were used in the preservation tests. Four of these were prepared at the base camp in the last few days of January. The remainder were made in the ship's refrigerator between February 10 and February 18. The thin sections were divided into two groups, the first of which was stored in the airtight can, the other in the case which was not airtight. A $1\frac{1}{2}$-inch cube of crevasse ice and another piece $1\frac{1}{2}$ inches square and approximately $\frac{1}{2}$ inch thick (pl. 2, *B*), were also stored in the airtight container. The thin sections of ice, as well as the larger pieces, were cemented to the glass slides by application of water under freezing conditions. An orientation symbol was scratched on the upper surface of both the $1\frac{1}{2}$-inch cube and the thinner fragment. The scratches were about $\frac{1}{16}$ inch deep.

In addition, irregular specimens of névé and crevasse ice were stored in individual airtight cans. Symbols were scratched on one or more faces of each specimen so that a rough estimate of the amount of wasting could be made.

The ice crystals (pl. 1, *B*), as well as all other specimens except the thin sections, were packed in snow in their containers, both to provide insulation and to absorb shocks encountered in transport. On arrival in Washington, the snow was easily brushed from the specimens.

### RESULTS

Of the 12 thin sections that had been prepared up to February 18, those that had been stored in the airtight container fared best. Of these, the sections prepared from the denser, less porous, and less pervious pressure-folded bay ice suffered the least wasting, showing only rounding of the corners. The $1\frac{1}{2}$-inch cube and the $\frac{1}{2}$-inch thick section were practically intact on arrival in Washington on April 14 (pl. 2, *B*). Such wasting as did occur in the thin sections and glass-mounted ice fragments probably took place on the numerous occasions when the specimens were exposed for study in the ship's refrigerator. This is suggested by the lack of wasting of the delicate ice crystals which were not exposed during the return trip. However, the ice crystals were packed in snow, the influence of which on the rate of wasting was not determined.

The sections stored in the slide box which was not airtight wasted rapidly. By March 24, 5–6 weeks after their preparation, only one was still a $1\frac{1}{2}$-inch square, although the corners were rounded and it had thinned appreciably. The remainder had wasted to remnants down to about $\frac{1}{2}$ inch in diameter. The well-preserved section, which may have been a little thicker than the others to start with, was transferred at this time to the airtight container. On arrival in Washington, 3 weeks later, its margins had contracted only about $\frac{1}{8}$ inch, in spite of the thinness of the section.

It is clear that preservation of thin sections of ice and transport over large distances are perfectly feasible if airtight containers are used and the storage temperatures are below freezing. Although the temperature of the ship's refrigerator was maintained at about 16° F., higher temperatures would probably not have appreciably affected the results of the tests.

A large part of the evaporation that did take place when the containers were open could probably have been eliminated by covering the thin sections of ice with cover glasses and sealing the edges

with vaseline, a procedure used by Perutz and Seligman (1939, p. 339). Unfortunately, the writer did not use such protected thin sections in his preservation tests. Warner (1945, p. 173) found that evaporation took place whether or not a covering was used, but he does not mention the steps taken to insure exclusion of air.

Although kerosene is an effective deterrent to evaporation, it has the disadvantage that, if frozen water is used as a cement and there are any chinks in the cement, the kerosene will be drawn into the section by capillarity. The large liquid bubbles beneath the section are particularly disturbing in petrographic examination. Eventually, the cement becomes weakened sufficiently so that the section comes loose. This situation is apparently aggravated by agitation of the kerosene during transport.

All irregularly shaped specimens were packed in snow in airtight cans. As pointed out earlier, markings were scratched to a depth of about $\frac{1}{16}$ inch on one or more surfaces so that a rough estimate of wasting could be made. In all cases, wasting was negligible, the markings having changed little if any on arrival in Washington, $2\frac{1}{2}$ months later. However, there was slight rounding of edges. Even this might have been prevented by application of vaseline or by soaking the snow in the containers with kerosene to eliminate air space. The use of vaseline or kerosene might also forestall possible sublimation growths.

Four delicate ice crystals (pl. 1, *B*) were also packed in snow in airtight containers. These crystals revealed no discernible evidence of wasting on examination in Washington. Unfortunately, in manipulating the crystal shown in plate 1, *B* for rephotographing, it broke into several pieces. Superficially, however, it seemed no different in Washington on April 14 than it did at Little America on January 28.

## SUMMARY AND CONCLUSIONS

Small ice specimens, including thin sections and delicate ice crystals, were brought back intact from the Antarctic, a distance of about 10,000 miles by the route traveled. On arrival in Washington, the specimens had been in storage for periods ranging from 2 to $2\frac{1}{2}$ months. Such long-distance transport with little or no wasting requires packing in airtight containers. Thin-section trays should be of the type in which the slides rest horizontally. The trays must rest in a framework which prevents play. Jostling of hand specimens of ice is easily prevented by packing in snow, which also serves as a good insulating medium. Storage in temperatures well below freezing is advisable, inasmuch as evaporation is proportional to the temperature. Unfortunately, experiments involving controlled temperatures were not possible on this expedition; hence no suggestions on optimum storage temperatures can be made. It is interesting to note, however, that well-stored and sheltered specimens showed no ill effects at Little America when the temperature rose to 28° F. The results obtained by Perutz and Seligman (1939) recommend the use of vaseline as a deterrent to evaporation. Kerosene has certain disadvantages, unless precautions are taken.

For ordinary petrographic studies, ice specimens no larger than hand specimens need be collected even from such distant places as Antarctica. Specimens of this size can be collected in relative quantity without exceeding the space or weight limitations under which Antarctic field

parties operate. Thus opportunity is afforded to sample the ice of the Antarctic with nearly the same degree of thoroughness as that with which bedrock is sampled.

The feasibility of transporting small specimens of ice over great distances opens up interesting possibilities. Heretofore, ice studies had had to be completed before expeditions left the Antarctic. This sometimes led to haste in completing studies. Now part or all of the studies involving loose samples of ice can be postponed until arrival at home. This has the added advantage that, if the collector is not himself qualified to undertake such studies, he can submit his specimens to specialists. In any event, research would benefit from the superior equipment and the better working conditions found in the laboratories at home.

If studies other than petrographic are contemplated, larger specimens of ice would probably be needed. This would reduce the number of specimens which field parties could collect. However, this limitation applies as well to rocks other than ice.

## REFERENCES CITED

DEMOREST, MAX (1940) The rock called ice: New York Acad. Sci. Trans., ser. 2, vol. 3, no. 2, pp. 25–28.

PERUTZ, M. F., and SELIGMAN, G. (1939) A crystallographic investigation of glacier structure and the mechanism of glacier flow: Royal Soc. London Proc., vol. 172, pp. 335–360.

WARNER, L. A. (1945) Summary of crystallographic investigations of the Ross Shelf ice in Repts. on Scientific Results of the U.S. Antarctic Service Expedition, 1939–1941: Am. Philos. Soc. Proc., vol. 89, pp. 172–17?.

# STUDIES FOR STUDENTS

## SECONDARY TILT: A REVIEW AND A NEW SOLUTION

KENNETH P. MCLAUGHLIN
State College of Washington, Pullman, Washington

### INTRODUCTION

The problem of secondary tilt is one familiar to most geologists who have worked in the field. The primary dip of cross-laminations may be intensified or dissipated as a result of tilting of the strata of which they are a part; strata underlying many angular unconformities are secondarily disturbed by postunconformity movement. Often it is desirable to know, in the field, the attitude of these and other primary and secondary structures prior to later disturbances to which they may have been subjected.

In this paper methods of solution of the "two-tilt" problem that have come to the writer's attention are very briefly reviewed. The new method of solution presented was devised in the course of field study involving cross-laminated beds.

### REVIEW OF SOLUTIONS

The earliest published solutions known to the writer are those of Harker (1884, pp. 154–162), who presented both graphic and algebraic methods. In the former, tangents are scaled as vector quantities, making visualization of the problem difficult.

Mertie (1922, pp. 572–574) referred the problem of secondary tilt to spherical trigonometry and showed how it may be solved by use of formulas based on spherical triangles. Fisher (1937, pp. 340–351) called attention to, and represented, Harker's original method of solution and later showed how the problem of two tilts may be solved by use of the stereographic projection (1938, pp. 1261–1271). Rapid preparation and interpretation of stereographic projections are made possible by use of a projection protractor devised by Fisher (1941, pp. 292–323, 419–442).

Bucher (1944, pp. 191–212) later presented the stereographic-net method of solution; in an appendix to the same paper, Fisher suggested a simplification of Bucher's method.

Spieker (1938, pp. 1255–1260) demonstrated that Harker's method is reliable only within certain narrow limits and that the illustration chosen by Fisher (1937) happened to fall within those limits. Spieker corrected Harker's original solution to make it applicable to all situations involving secondary tilt and presented another solution in which cotangents are plotted rather than tangents.

Corbett (1919, pp. 610–618) suggested a method by which contour maps may be used to work out original local structure in beds underlying an unconformity. By calculations involving the general (regional) strike and dip of rocks above and below the unconformity and the local structure of those above (with maps of both), the subsurface elevations of points on the strata as they were before secondary tilting may be obtained. With these as datum points, the original structure may be contoured. The method obviously has possibilities as applied to subsur-

face work but is somewhat unwieldy and impracticable for many stratigraphic and structural applications.

The various methods of solution referred to have certain disadvantages in that all require the use of data which may not always be available or make use of complicated formulas or involve map construction. However, for solution of problems involving numerous secondarily tilted units, such as the structural analyses of cross-lamination, the stereographic net is probably the best tool yet presented.

### SUGGESTED NEW METHOD OF SOLUTION

The solution here presented is completely graphic and makes use of only simple geometric principles. No materials are needed that are not normally a part of every field geologist's equipment.

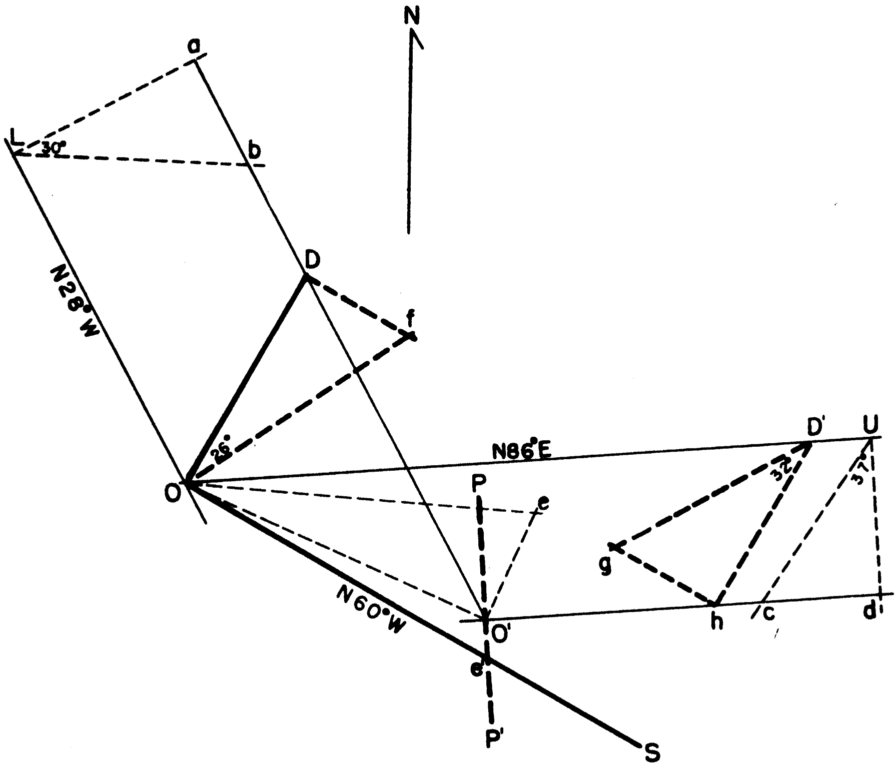



FIG. 1.—Graphic solution of a secondary-tilt problem. The fine-line construction is that used in determining the line of intersection of the two dipping planes. The construction in heavy lines is that by which the dip of *LOO'a*, prior to tilting of *UOO'd*, is determined.

Accuracy is dependent only on careful work and the scale used (fig. 1).

*Problem.*—Strata striking N. 28° W. and dipping 30° to the northeast are unconformably overlain by strata which strike N. 86° E. and dip 37° to the south. Determine the strike and dip of the un-

derlying strata prior to the deformation that tilted the overlying beds.

*Solution.*—With the method suggested by Reid (1909, p.174) and modified by Lahee (1941, p. 684), the strikes of the strata beneath and above the unconformity are represented by lines *LO* and *OU*, respectively; the angles of dip are represented by *aLb* and *cUd*. The difference in elevation of the lower beds between *L* and *a* (arbitrarily chosen) is *ab; cd* equals *ab* and is the difference in elevation of the upper beds between *U* and *d*. Thus *aO′* and *O′d* are the strike lines on a plane at *ab* (or *cd*) distance vertically below the plane of *LO* and *OU; LOO′a* and *UOO′d* are, therefore, planes representing the attitudes of the strata below and above the unconformity. Line *OO′* is the horizontal projection of the line of intersection of these two planes; *O′e* equals *ab* (and *cd*); and *O′Oe* is the angle of inclination of the line of intersection of the two planes.

If the entire construction is rotated about line *UO* until *UOO′d* is horizontal, line *Oe* (the actual line of intersection of the two planes) will also be horizontal and will coincide in direction with the strike of the underlying beds before they were secondarily tilted.

Line *PP′* is the trace of a vertical plane normal to *UO* and passing through *O′*. Before rotation, *e* is vertically below *O′* and in the plane *PP′*. Line *Oe′* equals *Oe* and occupies the new position of *Oe* after rotation; and *O′e′* is the projection in plan view of the arc through which *e* moves as rotation progresses. Thus *Oe′*, or *OS*, is the strike of the strata beneath the unconformity prior to the second tilting; and the dip direction is along line *OD*.

The original dip along *OD* is found by the use of the apparent dips of planes *LOO′a* and *UOO′d* along *OD*. Line *Df* equals *ab*, and the angle *DOf* is the present angle of apparent dip, along *OD*, of the strata beneath the unconformity. Similarly, the present apparent dip of the overlying strata in direction *DO* is the angle *gD′h*. As rotation progresses, the value of angle *gD′h* is added to that of *DOf*, and the sum of the two angles is the dip of the lower strata along *OD* prior to the second deformation. Thus the underlying strata (*LOO′a*) dipped 58° to the north with a strike of N. 60° W.

In problems in which the dip directions of the two sets of strata are in the same hemisphere, the dip of the older set prior to the last deformation is equal to the *difference between* the two apparent dips. Normally, the direction of dip—that is, whether from *O* to *D* or from *D* to *O* in figure 1—may be readily determined by inspection.

## REFERENCES CITED

BUCHER, W. H. (1944) The stereographic projection, a handy tool for the practical geologist: Jour. Geology, vol. 52, pp. 191–212.

CORBETT, C. S. (1919) Method of projecting structure through an angular unconformity: Econ. Geology, vol. 14, pp. 610–618.

FISHER, D. J. (1937) Some dip problems: Am. Assoc. Petroleum Geologists Bull. 21, pp. 340–351.

——— (1938) Problem of two tilts and the stereographic projection: Am. Assoc. Petroleum Geologists Bull. 22, pp. 1261–1271.

——— (1941) A new projection protractor: Jour. Geology, vol. 49, pp. 292–323, 419–442.

HARKER, ALFRED (1884) Graphical methods in field geology: Geol. Mag., new ser., decade 3, vol. 1, pp. 154–162.

LAHEE, F. H. (1941) Field geology, 4th ed., New York, McGraw-Hill Book Co.

MERTIE, J. B., JR. (1922) Analysis of structure below an unconformity: Econ. Geology, vol. 17, pp. 572–574.

REID, H. F. (1909) Geometry of faults: Geol. Soc. America Bull. 20.

SPIEKER, E. M. (1938) Problem of secondary tilt; Harker's solution corrected: Am. Assoc. Petroleum Geologists Bull. 22, pp. 1255–1260.

# GEOLOGICAL NOTES

## A NEW SPECIES OF EMBOLOMEROUS AMPHIBIAN FROM THE PERMIAN OF OKLAHOMA

J. WILLIS STOVALL
University of Oklahoma, Norman, Oklahoma

ABSTRACT

*Archeria victori*, sp. nov., is an amphibian from the Wellington (Lower Permian) collected near Eddy, Oklahoma. That the specimen represents an animal new to science is borne out by the character of the teeth and the absence of small secondary teeth on the lingual wall of the jaw.

### INTRODUCTION

The specimen described in this paper (U.O.M. 2-1-S8), was collected in the McCann rock quarry in the Wellington sandstone, 2 miles northeast of Eddy, Oklahoma. It consists of the posterior portion of the right mandible. Associated with it were numerous excellent skulls, and one nearly complete skeleton of *Captorhinus*, a few *Dimetrodon* bones, one *Trimerorhachis* skull, an *Eryops* left mandible, abundant abdominal scales of Archeria, and some indeterminate remains. All these forms are typically Permian and doubtless occur at a much later date than the *Cricotus* material described by Cope (1875–1876, pp. 404–411) from the Pennsylvanian of the Danville, Illinois, area. It seems likely that the Oklahoma specimen more nearly corresponds with the *Cricotus* material described by Cope (1878, pp. 505–530) from the Permian of northern Texas. Dr. A. S. Romer (1947, p. 269) has recently assigned the Texas cricotid forms to the genus *Archeria* on the ground that they are all from about the same horizon, that they are geologically much younger than the Illinois forms, and that "no adequate morphological distinctions are known."

The occurrence of all the forms mentioned above in the Briar Creek bone-bed in Archer County, Texas, suggests corresponding age. At that location Case (1915, p. 157) reported the following forms in association with "*Cricotus*":

*Dimetrodon incisivus* Cope
*Dimetrodon* sp., small form
*Theropleura* sp.
*Clepsydrops natalis* Cope
*Diadectes* sp.
*Diadectes maximus* Case
*Bolosaurus striatus* Cope
*Poliosaurus* sp.
*Archeria robinsoni* genus et sp. nov.
*Eryops megacephalus* Cope
*Cricotus heretoclitus* Cope
*Diplocaulus* sp.
*Trimerorhachis* (?) sp.
Dorsal plate of a member of *Dissorhophidae*
Dermal scutes of an amphibian
Numerous bones of amphibians and reptiles, many belonging to new forms, but some undoubtedly belonging to known forms described from incomplete material

Van Vleet (1901, p. 60) lists the following forms at Orlando:

*Diplocaulus magnicornis* Cope
*Diadectidae*, Gen. indt.
*Pariotichus incisivorus* (?) Cope
*Labyrinthodont*
*Trimerorhachis*

And, in 1927, Smith (1927, p. 37) reported the following forms from the same locality:

*Pleuracanthus cf. compressus*
*Gnathorhiza pusilla*
*Diplocaulus salamandroides*
*Trimerorhachis* sp.
*Crossotelos annulatus*

Romer (1935, p. 1625) reports the presence of *Archeria* in each formation of the Wichita group and in the Pueblo formation of the Cisco group in the Texas region. The genus has also been reported at Morrison (Smith, 1927, p. 42) (Wichita), Pond Creek (Smith, 1927, p. 4) (Clear Fork), and Orlando (Case, 1901, p. 63) (Clear Fork), Oklahoma, under the name of *Cricotus*. The writer has collected fragmentary remains of the genus in Wichita beds in Cotton, Stevens, and Jefferson counties, Oklahoma, where the formations are approximately the

equivalent of the Texas beds referred to by Romer. The Eddy find appears to extend the vertical range of this genus into the Clear Fork, although the position of the bone-bearing beds is 4 feet and 4 inches below the top of the Wellington, as shown by Clark and Cooper (1927, p. 10). Numbers 11 and 12 in the section given in table 1 are considered an outlier of the Garber.

The measured section described in table 1 is generalized, but the Wellington portion was taken on the side of the quarry from which the bones came.

TABLE 1*

MEASURED SECTION IN ROCK QUARRY IN NW¼ SEC. 16, T. 26 N., R. 2 W., KAY COUNTY, OKLAHOMA

| Material | Thickness | |
|---|---|---|
| | Feet | Inches |
| 13. Top soil | 2 | 0 |
| *Garber sandstone:* | | |
| 12. Sandstone, red, bedded, very fine-grained, ripple-marked; weathers into thin flagstones | 1 | 6 |
| 11. Sandstone, dark red, tightly cemented, hard | .... | 6 |
| *Wellington formation:* | | |
| 10. Clay-shale, gray, red and yellow banded | 1 | 2 |
| 9. Limestone, gray, very finely crystalline; grades locally into gray shale | .... | 0–4 |
| 8. Shale, dark gray, fissile | 1 | 4 |
| 7. Clay-shale, drab | 1 | 6 |
| 6. Sandstone, red with gray streaks, hard, with abundant small bone fragments; grades northward into gray silty shale (fossil horizon) | .... | 6 |
| 5. Sandstone, red, soft; grades northward into gray shale | .... | 6 |
| 4. Clay-shale, gray-green, soft | 3 | 0 |
| 3. Limestone, gray, very finely crystalline | .... | 0–1 |
| 2. Clay-shale, red | 4 | 0 |
| 1. Limestone, light gray, finely crystalline, with abundant vugs lined with calcite crystals | .... | 4 |
| | 16 | 6 |

* Measured by Ralph Disney and Lynn Jacobsen at the writer's request. This section is in the interfingered "zone of color change," where within a short distance the Wellington formation changes facies laterally southward from gray shale and mudstone with minor amounts of limestone to red shale and sandstone.

## DESCRIPTION OF MATERIAL

It is regrettable that the specimen is not complete. Nevertheless, the find is beautifully preserved, and it seems advisable to present descriptions at the present time with the hope that it will add to the knowledge of this poorly known form. Although only about two-thirds of the posterior portion of the right mandible was secured, the articular is complete and perfect, and the sutures are distinct. The complete length is approximately 345 mm., about the same size as the largest individuals of this genus discovered. The posterior half of the jaw (this specimen) of *Archeria* resembles both the pelycosaurs and the cotylosaurs, but, on the basis of the teeth, it is clearly of embolomerous affinities.

The height-width ratio of the specimen is distinctive, since it is relatively high and thin. The sides are nearly parallel. This specimen is, however, more obtuse in the angular region than any known species of this genus. The sutures are, in general, distinct, and the bones agree in shape only with the high-jawed amphibians.

In this specimen the articular, the prearticular, and the two coronoid bones are complete, while only a portion of the dentary, surangular, angular, and splenial are present. As in "*C.*" *heteroclitus*, there are two Meckelian fossae. The bones surrounding the large Meckelian fossa are distinct and complete, so that there is no doubt as to the size of the opening. The anterior end of the infra-Meckelian fossa can be only approximated because of the termination of the specimen at about that place. This opening is, however, long and narrow. Just anterior to and slightly above the infra-Meckelian fossa there appears to be the posterior end of a nutrient foramen. If this is the case, the present fossil corresponds in that particular to "*C.*" *heteroclitus*.

Posteriorly, the jaw is deep and quite obtuse—a feature that distinguishes it from all known species. The articular is complete and stands well above the tooth row.

The dentary of this specimen covers about one-fourth of the exterior surface of the jaw and is bounded below by the surangular. Posteriorly, it narrows toward the top of the jaw and is bounded above by the coronoid. As in the pelycosaurs and most amphibians, the thick upper surface of the dentary serves as attachment for

the lateral tooth row, which is bounded internally by the coronoid bones in this specimen.

The tooth row is not complete; but sixteen teeth are present, with gaps for three others and a stub of still another at the anterior end. The total number of teeth, therefore, cannot be determined; but the general form of the jaw suggests that there were between 45 and 52. There is no evidence of small teeth on the coronoid as in "*C.*" *heteroclitus*. The teeth preserved in this specimen are not differentiated as in most amphibians and are closely spaced, vertical, and remarkably uniform in size. The wear on the cutting edges of the teeth suggests a herbivorous diet, although Case (1911, p. 148) has suggested crocodilian habits for the beast. The teeth are decidedly chisel-shaped, as described by Case (1915, p. 160), and, in this respect, they resemble the marginal teeth of *Captorhinus*.

The teeth are peculiar in that they are flattened ovoid in cross section at the base, the greater diameter being perpendicular to the tooth row. This character is less pronounced in an anterior direction. Laterally, there is not much difference in diameter at the base and in the upper portion of the teeth. On the lingual

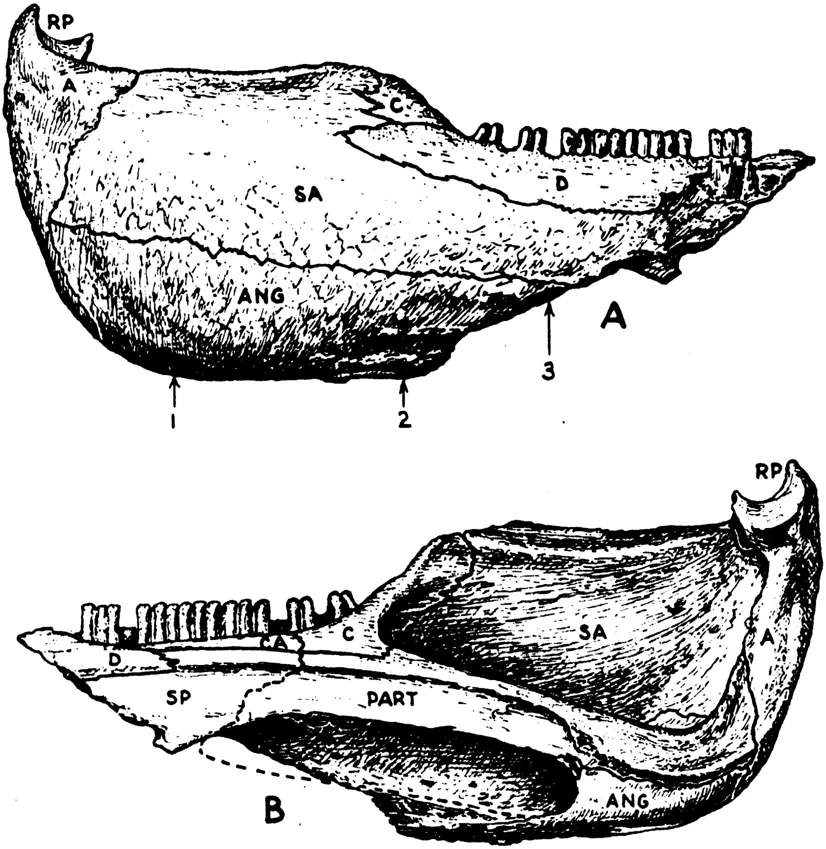



FIG. 1.—Two views of right mandible of *Archeria victori* sp. nov., from Eddy, Oklahoma. *A*, lateral view, length of fragment 192.2 mm., or about 7.5 inches. Figure is three-fifths natural size. *B*, internal view. *A*, articular; *ANG*, angular; *C*, coronoid; *CA*, anterior coronoid; *D*, dentary; *PRA*, prearticular; *RP*, retroarticular process; *SANC*, surangular; *SP*, splenial.

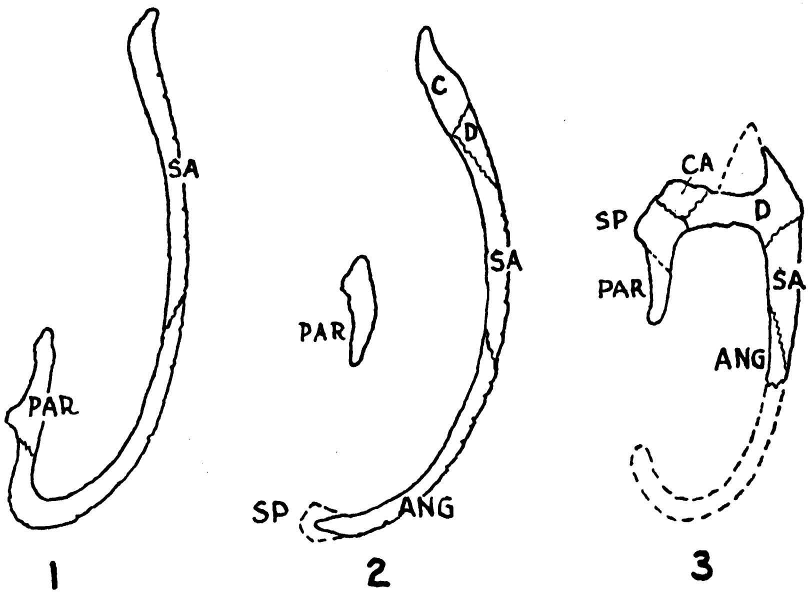



FIG. 2.—Sections through the jaw of *Archeria victori* sp. nov. at points *1, 2, 3*, indicated in *A* of Fig. 1. Each illustration natural size.

TABLE 2

MEASUREMENTS

| | Width (Mm.) | Length (Mm.) |
|---|---|---|
| Angular | 38.2 | 90* |
| Articular | 28.2 | 49.5 |
| Anterior coronoid | 4.4 | 38.9 |
| Coronoid | 18.8 | 48.1 |
| Dentary | 15.3 | 95 |
| Prearticular | 16.6 | 100.4 |
| Retroarticular process | 30.2 | ........ |
| Surangular | 48.4 | 141 |
| Splenial | 26.5 | 43.7 |

* Length of existing fragment.

side there are vertical striations that extend from the base to within 3 mm. of the top of the teeth. Similar striations occur on the teeth of some of the cotylosaurs and in all known species of "*Cricotus*."

When viewed from the side, the tops of the teeth form an almost straight line, and the length of the individual teeth ranges from 9.1 mm. at the posterior to 12 mm. at the anterior end. The width departs slightly from 3.6 mm. There is posterior overhang of the cutting edge on only a few of the teeth, a feature that is more pronounced in "*C.*" *heteroclitus* as described by Case (1915, p. 160).

There are two small coronoids, the true coronoid being visible above the posterior end of the dentary in an outside lateral view.

The angular forms the posteroventral portion of, and terminates posteriorly on, the outer surface of the ramus, continuing as a deep ventral keel on the posterior half of the jaw below the dentary and surangular. This bone is visible internally through the Meckelian fossa.

The articular forms the posterior margin of the ramus and is bounded externally by the surangular and angular and internally by the Meckelian fossa and the prearticular. The articular forms a margin or ventral keel between the inner and the outer surfaces of the jaw and extends upward, forming the retroarticular process, which in most other amphibians is well above the level of the tooth row.

The prearticular is a narrow bone, bounded posteriorly by the articular and angular and anteriorly by the coronoid and splenial. The pre-

articular also forms the lower boundary for the Meckelian fossa and serves as the upper margin of the infra-Meckelian fossa, forming the median wall of the posterior part of the Meckelian canal. This bone curves downward slightly and forms a bar across the inner surface of the jaw in a characteristic manner. Just anterior to the infra-Meckelian fossa, seen as a broad notch just below *SP* in *B* of figure 1, there appears to be an indication of a foramen in the splenial (see table 2).

## CONCLUSION

That this specimen is *Archeria* is suggested by the following characteristics: (1) the presence of like bones and similarity of their arrangement in the mandible; (2) the thin, deep character of the posterior portion of the jaw; (3) the arrangement of the teeth in a groove along the dentary; (4) the fact that the teeth are not differentiated in this portion of the jaw; (5) the position of the retroarticular notch and its resemblance to that in known specimens; (6) the parallel arrangement of the cutting edge of the teeth; and (7) the chisel shape of the teeth.

Specific characteristics of *A. victori* are: (1) the absence of small teeth on the coronoid; (2) the closer spacing of the teeth; and (3) the extreme bluntness of the posterior end of the jaws.

The species name, *victori*, is in honor of Professor Victor E. Monnett, who, as head of the Department of Geology at the University of Oklahoma, has been instrumental in the advancement of scientific research at the University.

## REFERENCES CITED

CASE, E. C. (1901) On some vertebrate fossils from Permian beds of Oklahoma: 2d Bienniel Rept. of Geology and Nat. History, Oklahoma Territory.

——— (1911) Revision of Amphibia and Pisces of the Permian of North America: Carnegie Inst. Washington Pub. 146.

——— (1915) Permo-Carboniferous red beds of North America and their vertebrate fauna: Carnegie Inst. Washington Pub. 207.

CLARK, G. C., and COOPER, C. L. (1927) Oil and gas geology of Kay, Grant, Garfield, and Noble counties: Oklahoma Geol. Survey Bull. 40-H.

COPE, E. D. (1875–1876) On fossil remains of Reptilia and fishes from Illinois: Nat. Acad. Sci. Proc., pp. 404–411.

——— (1878) Description of extinct Batrachia and Reptilia from the Permian formations of Texas. Am. Philos. Soc. Proc., vol. 17, pp. 505–530.

ROMER, A. S. (1935) Early history of Texas redbeds vertebrates: Geol. Soc. America Bull. 48.

——— (1947) Review of the Labyrinthodontia: Harvard Coll. Mus. Comp. Zoölogy Bull. 99, no. 1.

SMITH, G. N. (1927) The Permian vertebrates of Oklahoma, Thesis.

VAN VLEET, A. H. (1901) 2d Biennial Rept. of Dept. of Geology and Nat. History, Oklahoma Territory.

# AN EARLY REPORT OF ANCIENT LAKES IN THE BONNEVILLE BASIN

RONALD L. IVES
Indiana University

Most histories of geology credit Howard Stansbury with the discovery, in 1849, of field evidence indicating ancient higher lake levels in the Bonneville Basin (Stansbury, 1852; Gilbert, 1890, pp. 12–13). Stansbury's published reports indicate that he not only recognized the evidence for the ancient lakes but reached a reasonable conclusion as to their magnitude.

Recent translation and annotation, by Herbert S. Auerbach, of the diaries of the Escalante party, which traversed eastern and central Utah in 1776, makes available evidence which very strongly suggests that Stansbury's discovery was anticipated by nearly three-quarters of a century.

In the diary of Silvestre Veléz de Escalante, "a Franciscan Friar of great intelligence and ability," under the date of October 2, 1776, is the notation ". . . . This place, which we named Llano Salado, because we found some thin white shells there, seems to have once had a much larger lake than the present one." Location of this site, according to Auerbach's reconstruction of Escalante's route (Auerbach, 1943, pp. 1–142) is between Sevier Lake and the Beaver Mountains, in Millard County, Utah. Esca-

lante's diary includes a description (October 3, 1776) of areas in which is exposed "white, dry, fine loam, which from far away looked like white linen spread out to dry . . . ." (Auerbach, 1943, pp. 75–76; map opp. p. 90).

Aerial and jeep reconnaissances of the general region show that Auerbach's site identification is in close accord with the data recorded by Escalante but that a second site, east of the Beaver Mountains and south of Clear Lake, not far from Neels Station on the Los Angeles and Salt Lake Railroad, fits the description equally well. At both sites, "white, dry, fine loam . . . ." (Gilbert's "white marl"), containing large numbers of shells (largely *Lymnaea bonnevillensis*), is present.

In view of the foregoing, it appears that Escalante discovered and reported evidence of high lake levels in the Bonneville Basin some seventy-three years before Stansbury. Present historical knowledge indicates that this is the earliest such report, as the Escalante expedition was the first official party to enter the Utah Desert.

## REFERENCES CITED

AUERBACH, H. S. (1943) Father Escalante's journal, with related documents and maps: Utah State Hist. Quart., vol. 2.

GILBERT, G. K. (1890) Lake Bonneville: U.S. Geol. Survey Mon. 1.

STANSBURY, HOWARD (1852) Exploration and survey of the valley of the Great Salt Lake of Utah. Philadelphia.

# THE COMPOUND $CaO \cdot 2Al_2O_3$

JULIAN R. GOLDSMITH
University of Chicago

In the original work on the system $CaO$–$Al_2O_3$ (Shepherd, Rankin, and Wright, 1909, pp. 293–333) a compound $3CaO \cdot 5Al_2O_3$ was identified as one of the calcium aluminates stable at liquidus temperatures. The composition was determined by microscopic examination of synthetically prepared mixtures, fused in an iridium furnace. Two modifications of this compound were described—a stable, uniaxial modification, with $\epsilon = 1.651$ and $\omega = 1.617$, and an "unstable" biaxial form, with $\alpha = 1.662$, $\beta = 1.671$, $\gamma = 1.674$, and a $2V$ of $35° \pm 5°$. Individual crystals of the unstable modification were not observed; they all showed evidence of alteration to the uniaxial form, which proceeded inward from the periphery of the grains.

In 1937 two papers were published within a few weeks of each other, one by Lagerqvist, Wallmark, and Westgren (1937, pp. 1–16) working in Sweden, the other by Tavasci (1937, pp. 717–719) in Italy, in which the conclusion was independently reached that the $3CaO \cdot 5Al_2O_3$ formulation was incorrect, the ratio of lime to alumina in reality being 1:2. The conclusions of Lagerqvist *et al.* were based on X-ray evidence in conjunction with density determinations, as well as powder photographs. It should be noted that the Weisenberg determinations showed the material to be monoclinic, with $a = 18.82$ A, $b = 8.84$ A, $c = 5.42$ A, and $\beta = 107°.8$. No optical determinations were made. Tavasci arrived at the $CaO \cdot 2Al_2O_3$ composition on the basis of examination of etched polished specimens with the ore-microscope. He found that crystallized melts of the 3:5 ratio were inhomogeneous, whereas a homogeneous single crystalline phase developed from a melt of the composition $CaO \cdot 2Al_2O_3$. This compound melted incongruently at a temperature of about 1765° C. No data on optical properties are given in Tavasci's paper, but he states that they are the same as those assigned by Wright to the 3:5 compound.

The American literature has virtually ignored this 1937 work; the only references containing any mention of even the possibility of the existence of the 1:2 compound known to the writer are (*a*) the A.S.T.M. X-ray diffraction data cards and (*b*) Bulletin 107 of the National Research Council: *Data on Chemicals for Ceramic Use.* Both these sources are merely compilations of data, and in neither is the question of composition clarified; the reader cannot be certain as to which formulation is correct or whether both compounds exist. The 1942 printing of the phase diagram for the system $CaO$–$Al_2O_3$–$SiO_2$ (Sosman and Anderson) shows the

compound $3CaO \cdot 5Al_2O_3$ as the calcium-aluminate in question, as does the diagram just published in the compilation of phase diagrams by the American Ceramic Society (Hall and Insley, 1947). The fact that the American literature is not up to date on this subject led the writer, in a discussion of solid-solution relations in a recently published paper (1947, pp. 381–404), to mention "a hitherto unknown 'molecule,' of the composition $CaO \cdot 2Al_2O_3$."

Some experimental work was done to help clear up the problem and as a check on the data of Tavasci and of Lagerqvist, Wallmark, and Westgren. A mixture of CaO (as $CaCO_3$) and $Al_2O_3$ in the 1:2 ratio was carefully prepared; it was thoroughly mixed, sintered at approximately 1500° C., crushed, remixed, and resintered. Portions of the mass were then fused in an oxygen-gas flame. Microscopic examination confirmed Tavasci's observation—a single homogeneous phase had crystallized from the melt. It is interesting to note that virtually complete reaction had taken place during sintering at 1500° C. in a relatively few hours, as seen microscopically. The optical properties are indeed those described by Wright (1909); the crystals are uniaxial (+), although, as Wright observed, some show a slight tendency toward biaxial development. The indices of refraction are $\omega = 1.617 \pm 0.002$ and $\epsilon = 1.651 \pm 0.002$ (sodium vapor light). No trace of the "unstable phase" was observed, although no special effort was made to obtain it. X-ray powder pictures show the same pattern as that obtained by Lagerqvist and co-workers. Their technique apparently did not permit the observation of diffraction lines of low $\theta$ value, however, as several lines are present that they did not find; one strong line of $d = 4.44$ (indices 020) and a moderately weak line of $d = 6.17$ (indices 200) are worth mentioning. Slight discrepancies in spacings might indicate that some refinement of unit-cell dimensions is desirable. No crystals sufficiently good for a Weisenberg determination were obtained.

The 1909 diagram of Rankin and Shepherd indicated that the 3:5 compound melted incongruently. In 1915 a revised diagram by Rankin and Wright (p. 11) has the compound as a congruent one, possibly as a result of the location of the boundary curve in the ternary system. The 1:2 ratio of lime to alumina places the composition point outside its stability field, however, if the boundary curve is correctly placed, and thus, as pointed out by Tavasci, $CaO \cdot 2Al_2O_3$ melts incongruently.

No additional information has been obtained as to the existence of an unstable modification. The X-ray diffraction powder picture obtained by Lagerqvist, Wallmark and Westgren indicates that they dealt with the same modification here described, in which case the compound is monoclinic. Observation of optical properties alone would lead one to the conclusion that $CaO \cdot 2Al_2O_3$ is hexagonal or tetragonal.

ACKNOWLEDGMENT.—This work was done in conjunction with a research program made possible by financial assistance from the Office of Naval Research.

## REFERENCES CITED

GOLDSMITH, J. R. (1947) The system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$: Jour Geology, vol. 55, pp. 381–404.

HALL, F. P., and INSLEY, HERBERT. (1947) Phase diagrams for ceramists: Am. Ceramic Soc. Jour., vol. 30, pt. 2.

LAGERQVIST, VON KARIN; WALLMARK, SIGNE; and WESTGREN, A. (1937) Röntgenuntersuchung der System $CaO$–$Al_2O_3$ und $SrO$–$Al_2O_3$: Zeitschr. anorg. Chemie, vol. 234, pp. 1–16.

RANKIN, G. A., and WRIGHT, F. E. (1915) The ternary system $CaO$–$Al_2O_3$–$SiO_2$: Am. Jour. Sci. 4th ser., vol. 39.

SHEPHERD, E. S.; RANKIN, G. A.; and WRIGHT, F. E. (1909) The binary systems of alumina with silica, lime and magnesia: Am. Jour. Sci., 4th ser., vol. 28, pp. 293–333.

SOSMAN, R. B., and ANDERSON, OLAF. (1942) Phase diagram for the system $CaO$–$Al_2O_3$–$SiO_2$, U.S. Steel Corp. Reprinted with corrections from the 1933 diagram.

TAVASCI, B. (1937) Tonindustrie-Zeitung, vol. 61, pp. 717–719, 729–73.

# COMMUNICATIONS TO EDITOR

## TRANSLATIONS OF RUSSIAN GEOLOGIC LITERATURE

The Committee on Russian Literature of the Geological Society of America is compiling a list of English translations of Russian geologic papers and books covering all fields from geophysics to paleontology. Information regarding manuscript translations is desired. It should be sent to the chairman of the committee, Ronald K. DeFord, Box 1814, Midland, Texas, or to the secretary of the Geological Society. Such information will be published in the forthcoming volumes of the *Bibliography and Index of Geology Exclusive of North America* and the *Bibliography of Economic Geology.*

## PUBLICATION OF THE GEOPHYSICAL ABSTRACTS

The United States Geological Survey has resumed publication of the *Geophysical Abstracts* after a four-year interval, during which they were issued by the United States Bureau of Mines.

The *Geophysical Abstracts* are published quarterly as an aid to those engaged in geophysical research and exploration. The bulletin covers world literature on geophysics contained in periodicals, books, and patents. It deals with exploration by gravitational, magnetic, seismic, electrical, radioactive, geothermal, and geochemical methods and with underlying geophysical theory and related subjects.

Copies may be purchased singly or by annual subscription from the Superintendent of Documents, Government Printing Office, Washington 25, D.C. For subscription, the Superintendent of Documents will accept a deposit of $5.00 in payment for subsequent issues. When this fund is near depletion, the subscriber will be notified. The deposit may also be used to cover purchase of any other publication from the Superintendent of Documents.

# REVIEWS

"Pollen Profile from a Texas Bog." By J. E. Potzger and B. C. Tharp. (*Ecology*, vol. 28.) Pp. 274–280, 1947.

The pollen profile is from the Patschke Bog, which is located 40 miles east of Austin in eastern Texas. The bog is remarkable for its great depth—22 feet—considering the low latitude, and for containing pollen grains of *Picea glauca* (=*canadensis*), *P. mariana*, and *Abies*. Samples were taken at 1-foot intervals. The significant vegetational changes and the climatic conclusions are briefly discussed below.

*Foot-levels 22–18* (*at bottom of bog*).—Maxima of *P. glauca*, *P. mariana*, and *Abies* indicate a cool climate. Increase of predominantly medium-sized grass pollens from a minimum at the bottom to a maximum at foot-level 18 suggests a change from moist to dry climate. However, tall-grass prairie species, which have large pollen grains, did not predominate at these or at any other levels in the bog.

*Foot-levels 17–16.*—Maxima of medium-sized grass pollens and of *Quercus* and absence or scarcity of boreal conifers suggest a warm-dry climate.

*Foot-levels 15–10.*—Abundance of *Alnus*, moderate representation of *Castanea*, and minimum of medium-sized grass pollens suggest a warm-moist climate. The present western range limit of *Castanea* is about 100 miles to the east of the Patschke Bog, where the rainfall is several inches greater.

*Foot-levels 9–6.*—Maximum of *Castanea* and abundance of small-sized grass pollens (moist-habitat forms) suggest a continued warm-moist climate.

*Foot-levels 5–1* (*at top*).—Maximum of medium-sized grass pollens, increase of *Quercus*, appearance of *Carya*, and decline and disappearance of *Castanea* indicate a warm-dry age.

The present vegetation of the region consists of the oak-hickory association. Tall-grass prairie begins 15 miles to the west. The climate is moist subhumid, and mesothermal (Thornthwaite, 1941, pl. 3).

The climatic history is thus briefly as follows:

| Foot-Levels | |
|---|---|
| 0 = top | |
| | Warm-dry |
| 5.5 | |
| | Warm-moist |
| 15.5 | |
| | Warm-dry |
| 17.5 | |
| | Cool, moist, growing dry |
| 22 | |

*Picea glauca* and *P. mariana* are boreal trees whose modern southern range limit skirts the southern tip of Lake Huron and passes through central Wisconsin and the northwestern corner of Minnesota, following in this region the July isotherm of 70° F. (U.S. Dept. Agr., 1941). The pollens of these spruces in the Patschke Bog thus occur over 1,000 miles, or 13 degrees of latitude, south of the modern ranges of the trees. Cones, wood, and twigs of *P. glauca* in Pleistocene beds at Percy Bluff, 35 miles north of Baton Rouge in Louisiana, show that this spruce once really grew in these low latitudes (Brown, 1938, pp. 67, 86). Since the present-day July temperature at the Patschke Bog is 84° F., the July temperature in the region then was perhaps some 14° F. lower than now. A Pleistocene lowering of the July temperature of 13°–14° F. (7°.28–7°.8 C.) is also suggested by pollens of *Picea* and *Abies* in Florida (Davis, 1946); by pollens of a vegetation strikingly resembling the modern transition forest of northern Minnesota at the bottom of the Jerome Bog northwest of Wilmington in North Carolina (Buell, 1946, pp. 14–15); and by fossil marmots 4,000 feet above their modern level of range in New Mexico (Stearns, 1942, pp. 867–878).

The relative thinness of the cool bottom beds in the Patschke Bog suggests that they hardly extend beyond the last, the Mankato, glacial maximum, making the whole bog represent the last twenty-five thousand years.

The climatic history inferred by Potzger and Tharp resembles that concluded by Paul Sears

and E. S. Deevey for the "Lateglacial" and the "Postglacial" of the northeastern United States, except for the absence of a relatively cool and moist age, Sears and Deevey's "Subatlantic," during the past few thousand years. Of course, a correlation requires that Sears and Deevey's time scale be stretched from ten thousand to some thirty thousand years.

The climatic history of the bog shows less agreement with that of the region extending from central Texas into the state of Washington. In this vast area the climate became increasingly warm and dry until a few thousand years before Christ; and since about 2000 B.C. it has been predominantly moderately cool and moist but with dry (and warm?) periods. This seems to be the proper characterization of the last stage in Arizona and western Texas, for the recent erosion-recorded droughts can have been but brief interludes, to judge from the rapidity of the modern erosion in the region. One drought, recognized in northern (Hack, 1942, pp. 58, 68) and southeastern (Sayles and Antevs, 1941, pp. 43, 46) Arizona, in Trans-Pecos Texas (post-Calamity erosion) (Bryan and Albritton, 1943, pp. 486–487), and on the Texas High Plains ("renewed deflation") (Evans and Meade, 1945, pp. 499–503) occurred perhaps A.D. 1276–1299 (1273–1300), when narrow tree-rings were formed on the Colorado Plateau (Schulman, 1947, pp. 2–8). Another drought, recorded by erosion in southeastern Arizona and probably by an unconformity in the Calamity formation in Trans-Pecos Texas, seems to have taken place early in the Christian era. Consequently, the now moist and moist subhumid eastern Texas and the semiarid and arid country from western Texas to Washington seem to have had opposite temperature trends during the last four thousand years and opposite moisture trends almost throughout the Postpluvial.

## REFERENCES CITED

Brown, C. A. (1938) The flora of Pleistocene deposits in Louisiana: Louisiana Dept. of Cons. Geol. Bull. 12, pp. 59–96.

Bryan, Kirk, and Albritton, C. C. (1943) Soil phenomena as evidence of climatic changes: Am. Jour. Sci., vol. 241, pp. 469–490.

Buell, M. F. (1946) The age of Jerome bog, "a Carolina bay": Science, vol. 103, pp. 14–15.

Davis, John, Jr. (1946) The peat deposits of Florida: Florida Geol. Survey Bull. 30.

Evans, G. L., and Meade, G. E. (1945) Quaternary of the Texas high plains: Univ. Texas Pub. 4401, pp. 485–507.

Hack, J. T. (1942) The changing physical environment of the Hopi Indians of Arizona: Peabody Mus. Papers, Harvard Univ., vol. 35, no. 1.

Sayles, E. B., and Antevs, Ernst (1941) The Cochise culture: Medallion Papers no. 29.

Schulman, Edmund (1947) An 800-year Douglas fir at Mesa Verde: Tree-Ring Bull. 14, pp. 2–8.

Stearns, C. E. (1942) A fossil marmot from New Mexico and its climatic significance: Am. Jour. Sci., vol. 240, pp. 867–878.

Thornthwaite, C. W. (1941) Atlas of climatic types in the United States, 1900–1939: U.S. Soil Cons. Service Misc. Pub. 421.

U.S. Department of Agriculture (1941) Climate and man: Yearbook for 1941.

Ernst Antevs

---

*Elements of Soil Conservation*. By Hugh Hammond Bennett. 1st ed. New York: McGraw-Hill Book Co., Inc., 1947. Pp. x+406; figs. 114. $3.20.

The broader problems of soil wastage and methods of soil conservation are here effectively presented in nontechnical terms. About one-third of the book is devoted to erosion processes and their effects and the remaining two-thirds to standard methods of soil conservation. Questions for discussion are given at the end of each chapter, and at the end of the book there is a list of films and film strips which can be used to supplement the text. The book is suitable as an elementary text for both student and adult groups. Teachers of beginning geology may find it useful for supplementary readings. It does not in any sense replace the same author's *Soil Conservation* as a comprehensive reference book.

L. H.

THIS issue contains the concluding article in the series on the "Composition of Meteoritic Matter," by Harrison Brown and Claire Patterson. An oral presentation of this research at the recent Chicago sessions of the American Association for the Advancement of Science won the $1,000 prize for the outstanding paper given at the meeting. The implications regarding the origin of meteorites and of the earth's interior will be of interest to every geologist and astronomer.

A forthcoming number of the *Journal* will contain papers presented at the A.A.A.S. as a "Symposium on Problems of Mississippian Stratigraphy and Correlation."

Among the articles to appear in other early issues are:

"A Preliminary Report of Vertebrates from the Permian Vale Formation of Texas." By EVERETT C. OLSON

"Cave-in Lakes in the Nabesna, Chisana, and Tanana River Valleys, Eastern Alaska." By ROBERT E. WALLACE

"Hollow Ferruginous Concretions in South Carolina." By L. L. SMITH

"Geological Significance of Surface Tension." By JEAN VERHOOGEN

"Oxidation and Reduction in Geochemistry." By BRIAN MASON

"A Note on the Original Isotopic Composition of Terrestrial Carbon." By KALERVO RANKAMA

## SUGGESTIONS TO CONTRIBUTORS

TEXT—Manuscripts should be written in English, or in French or German with an adequate summary in English. They should be typewritten, on one side of paper only, double spaced, and with wide margins.

ABSTRACT—Each article should contain an abstract which gives a concise summary of the content, including both major and minor points. Abstracts should not exceed 200 words.

REFERENCES—Under the heading "References Cited" at the end of the paper should be listed all literature cited, arranged alphabetically by authors and chronologically under each author, using the following form:

DOE, J. S. (1884*a*) Geology of the Big Ben Mountains: U.S. Geol. Survey Prof. Paper 55, pp. 7–17, pl. 2.
——— (1884*b*) Guidebook for excursions in the Big Ben Mountains, pp. 32, 33, New York, Macmillan Co.
SMITH, A. V. (1926) Discovery of fish remains in Ordovician rocks of the Black Hills (abstr.): Geol. Soc. America Bull. 52, p. 27.
———, and DUNSANY, A. J. (1929*a*) Early Ordovician faunas of South Dakota: U.S. Geol. Survey Bull. 444, pp. 1–76, 3 pls.
———, ——— (1929*b*) Revision of Ordovician stratigraphy of the Black Hills: Jour. Geology, vol. 37, pp. 680–695.

In the text, references to the literature cited should follow this form: (Smith, 1928, p. 36).

FOOTNOTES—Discussion which has a collateral bearing on the subject being discussed but would be too much of an interruption to be put in the body of the text should be inserted as a footnote. Footnotes should be inserted immediately after references are made to them, with lines above and below setting them off from the text. They should be numbered consecutively.

ABBREVIATIONS—This *Journal* uses the abbreviations approved by the United States Geological Survey and listed in the Survey's *Suggestions to Authors.*

ILLUSTRATIONS—Maps, line drawings, and diagrams should be prepared on white drawing paper or on tracing cloth, with India ink. Care should be taken not to overload maps with irrelevant detail or, at the other extreme, to use excessive amounts of space to convey relatively little information. The smallest symbols or letters used should be of such size that they will be not *less* than 1 mm. high after reduction. Unsatisfactory illustrations will be returned to the author or may, with his permission, be redrafted at the editorial office at his expense.

Photographs are reproduced by the collotype process or as halftones. Authors should use the minimum number consistent with adequate presentation of the subject. If a paper includes a disproportionately large number of plates, the editor may ask the author to pay a portion of the cost of illustrations. Photographs should be sharp and clear, printed on glossy paper, and either of the size in which they are to appear in the *Journal* or larger.

Explanations of all figures should be typed on one or more separate sheets. Each figure should be marked for identification and top indicated if there is any doubt.

The original illustrations are destroyed following their publication, unless the author has requested their return in advance of publication.

ALTERATIONS—The cost of excessive alterations (i.e., changes from the original manuscript) made by the author will be charged to him. No changes should be made in proof except errors of typography or of fact based on new information.

REPRINTS—The *Journal* supplies free 50 reprints (without covers) of each major article. This does not include geological notes, discussions, reviews, or communications to the editor. Additional reprints may be ordered when manuscript is submitted or at any time in advance of publication. Rates will be furnished on request.

STYLE—In matters of style, spelling, abbreviation, etc., the two guides to be followed are the University of Chicago Press *Manual of Style* and the United States Department of the Interior *Suggestions to Authors.* Where these authorities differ, the latter reference is to be used.

VOLUME 56 NUMBER 2

# THE JOURNAL OF GEOLOGY

*March 1948*

## THE COMPOSITION OF METEORITIC MATTER

### III. PHASE EQUILIBRIA, GENETIC RELATIONSHIPS AND PLANET STRUCTURE

HARRISON BROWN AND CLAIRE PATTERSON
Institute for Nuclear Studies, University of Chicago

ABSTRACT

An attempt has been made to study the composition of meteorites on a more quantitative basis than has been attempted heretofore. In particular, investigations have been made of experimental and theoretical approaches that might lead to more rigid comparisons between terrestrial and meteoritic matter. Stress has been placed particularly upon the following studies: (*a*) the distribution of elements between meteoritic phases; (*b*) average composition as a function of metal-phase content; (*c*) correlation between element distribution, thermochemical data, and general thermodynamic considerations.

It is demonstrated that if one assumes that the observed distributions of elements represent equilibrium distributions, then equilibrium must have been established at temperatures of the order of 3000° C. and pressures of the order of $10^5$–$10^6$ atm. Similarly, it is demonstrated that the conditions at which equilibrium was achieved varied from meteorite to meteorite in such a way that the greater the metal-phase content, the greater the temperature and/or the pressure. The data indicate strongly that meteorites had their origin in a planet similar to the earth in general physicochemical characteristics.

#### INTRODUCTION

In 1850 A. Boisse suggested that the earth's composition is probably equivalent to the mean composition of meteoritic matter. Since that time much effort has been expended by astronomers, geologists, geophysicists, and, more recently, geochemists in attempts either to develop or to disprove Boisse's original speculation. On the whole, information accumulated during the last fifty years, notably by such men as Farrington, Merrill, Prior, Tammann, Goldschmidt, and Jeffreys, has served to substantiate in a qualitative way the thesis that meteorites belong to a single family possessing a common genesis, in all likelihood a planet similar in physicochemical characteristics to the earth. In particular, the investigations to date have resulted in the following conceptual advances: (1) belief that the dense core of the earth, as defined by seismic data, is probably similar to iron-nickel meteorites in composition; (2) belief that the silicate phase of stony meteorites is similar to the mantle surrounding the core of the earth; (3) realization that chemical equilibrium, or at least something approaching equilibrium, at one time existed between the various meteoritic phases; and (4) realization, in a qualitative way, of the physicochemical principles underlying the differentiation of elements during magma crystallization and resulting in the formation of igneous rocks from a mother-substance not too dissimilar in composition to the silicate phase of stony meteorites.

It must be emphasized that these

general comparisons have arisen almost entirely in a qualitative way, gross averages of meteoritic compositions having been used, in discussions both of the differentiation of elements during magma crystallization and of phase equilibria. More quantitative comparisons are highly desirable if the nature of the relationships, if any, between the earth and meteorites is to be ascertained and, looking still farther ahead, if the physicochemical processes that took place during the process of planet formation are to be understood, even partially.

This paper represents an attempt to study existing meteoritic data from a somewhat more quantitative point of view and to investigate experimental and theoretical approaches that might lead to more rigid comparisons between terrestrial and meteoritic matter.

## THE AVERAGE COMPOSITION OF METEORITIC MATTER

The average composition of meteoritic matter tells us little about meteorites as a family. However, during the course of the present discussion, average compositions will be referred to from time to time; so, for convenience, the average compositions of the silicate phase of stony meteorites and iron meteorites and the gross composition of stony meteorites are given in table 1. The data given in columns 1, 2, and 3 are taken from recent recomputations by the authors (1947*a*, pp. 405–411; 1947*b*, pp. 508–510), and column 4 has been derived by combining the data of columns 1 and 2 with new computations of the average metal and sulphur contents of stony meteorites. The latter two averages are given in table 2, together with their standard deviations and precisions. Sulphur has been calculated as displacing oxygen when unaccounted for as FeS. The combined iron, shown in column 4 of table 1, represents the iron combined as oxide, sulphide, and phosphide.

The errors associated with the various components listed in column 4 of table 1 are greater, relative to one another, than in the other columns, owing largely to the substantial size of the standard deviation associated with the metal phase. The reasons for this unusually large standard deviation will become apparent when the metal-phase frequency curve is discussed later in the paper.

## PHASE EQUILIBRIA IN METEORITES

Perhaps the most convincing evidence that fragments of meteoritic matter, now widely scattered, at one time existed in close contact with one another is the evidence derived from the distribution of elements between the various meteoritic phases. There are three main phases: the silicate phase, composed primarily of $SiO_2$, MgO, and FeO; the metal phase, composed primarily of iron, nickel, and cobalt; and troilite, composed primarily of FeS. In 1910 W. A. Wahl (p. 67) pointed out that those major meteoritic constituents whose oxides possess low heats of formation exist primarily in the metal phase and that those constituents whose oxides possess high heats of formation exist primarily in the silicate phase. The comprehensive studies by I. and W. Noddack (1930, p. 757; 1934, p. 173), V. M. Goldschmidt (1935, p. 183), and others on the distribution of minor constituents between the silicate and metal phases have well substantiated the fact that the phase (metal or silicate) in which a given element tends to concentrate depends markedly upon the affinity of the element for oxygen.

Prior (1916, p. 26), surveying meteoritic compositions more closely, observed that, in general, the smaller the

concentration of metal phase in a stony meteorite, the greater the concentration of nickel in the metal phase and the greater the concentration of combined iron in the silicate phase. Prior (1920, p. 54) advanced the explanation that "meteorites have separated from a single magma which has passed through successive stages of progressive oxidation."

The Ni–$FeSiO_3$–$NiSiO_3$–Fe equilibrium has been studied by Zur Strassen (1930, p. 209), who measured the equilibrium constant for the reaction

$$Ni + FeSiO_3 \rightleftharpoons Fe + NiSiO_3 \ ,$$

where the constant $C$ is given by the expression

$$C = \frac{(Fe)_m (Ni)_{si}}{(Ni)_m (Fe)_{si}} .$$

Zur Strassen found that the reaction obeys the mass-action law over a wide range of iron and nickel compositions and that, in the absence of sulphur, the equilibrium constant is independent of the composition of the silicate phase. He found (1931, p. 46) that sulphur slightly displaces the equilibrium. The constants as measured by Zur Strassen at two temperatures in the absence of sulphur are given in table 3.

The observed "average" equilibrium constant for iron and nickel in stony meteorites as calculated from table 1 is

$$C = 0.24 \ ,$$

a value nearly forty times greater than

TABLE 1

THE AVERAGE COMPOSITION OF STONY AND IRON METEORITES
(Per Cent by Weight)

| Element | 1 Silicate Phase of Stones | 2 Metal Phase of Stones | 3 Iron Meteorites | 4 Total Stony Meteorites |
|---|---|---|---|---|
| Oxygen | 43.12±1.00 | | | 36.15±0.89 |
| Silicon | 21.61±0.19 | | | 18.12±0.22 |
| Comb. iron | 13.23±0.42 | | | 14.21±0.39 |
| Comb. nickel | 0.39±0.07 | | | 0.33±0.06 |
| Comb. cobalt | 0.02±0.01 | | | 0.02±0.01 |
| Magnesium | 16.62±0.32 | | | 13.93±0.29 |
| Sulphur | | | | 1.79±0.08 |
| Calcium | 2.07±0.21 | | | 1.74±0.18 |
| Aluminum | 1.83±0.15 | | | 1.53±0.13 |
| Sodium | 0.82±0.06 | | | 0.69±0.05 |
| Chromium | 0.36±0.06 | | | 0.30±0.05 |
| Manganese | 0.31±0.07 | | | 0.26±0.06 |
| Potassium | 0.21±0.02 | | | 0.18±0.02 |
| Phosphorus | 0.17±0.01 | | | 0.14±0.01 |
| Titanium | 0.10±0.03 | | | 0.08±0.03 |
| Hydrogen | 0.07±0.02 | | | 0.06±0.02 |
| Metallic iron | | 88.58±0.55 | 90.78±0.26 | 9.97±0.69 |
| Metallic nickel | | 10.68±0.51 | 8.59±0.24 | 1.20±0.10 |
| Metallic cobalt | | 0.70±0.06 | 0.63±0.02 | 0.08±0.01 |

TABLE 2

CONCENTRATION OF METAL PHASE AND SULPHUR IN STONY METEORITES

| Substance | Average Concentration (Per Cent) | Standard Deviation (Per Cent) | Number of Cases | Precision (Per Cent) |
|---|---|---|---|---|
| Metal phase | 11.25 | 8.94 | 130 | 0.78 |
| Sulphur | 1.79 | 0.93 | 130 | 0.08 |

Zur Strassen's measured value at 1750° K. A study of the effects of sulphur serves to demonstrate that the average sulphur content of stony meteorites (1.79 per cent by weight) is too small to produce any appreciable effect upon the equilibrium under the conditions used by Zur Strassen. One is thus forced to the conclusion that, if equilibrium existed between the metal and silicate phases of stony meteorites, it must have been established under conditions of temperature or pressure markedly different from 1750° K. and 1 atm.

TABLE 3

MEASURED EQUILIBRIUM CONSTANTS FOR THE REACTION $Ni + FeSiO_3 \rightleftarrows NiSiO_3 + Fe$
(Data of Zur Strassen [1930])

| Temperature ° K. | $C$ | Remarks |
|---|---|---|
| 1750 | $6.53 \times 10^{-3}$ | Average |
| 1840 | $7.25 \times 10^{-3}$ | Average |
| 1840 | $7.44 \times 10^{-3}$ | Highest value |

That the temperature effect is in a direction indicating that equilibrium took place at a temperature higher than 1750° K. can be seen from the fact that the reaction is endothermic and by applying the relationship[1]

$$\frac{d\ln K}{dT} = \frac{\Delta H}{RT^2} \quad (\Delta H = \text{The enthalpy change for the reaction}),$$

thus obtaining a positive value for the change of $\ln K$ with increasing temperature. However, preliminary calculations, to be presented at a later date, indicate that the high observed value of the "average" equilibrium constant in meteorites cannot be easily accounted for on an increased-temperature basis alone, the temperature obtained being unreasonably high.

That increasing pressures would tend to push the equilibirum concentration toward the nickel silicate side can be seen from an inspection of the relative volumes of iron and nickel metals and oxides, indicating that a volume decrease probably results from the silicate reaction. The change in volume is illustrated in table 4.[2]

TABLE 4

VOLUME CHANGE IN THE REACTION
$Ni + FeO \rightarrow NiO + Fe$

| Substance | Molar Volume (cc.) |
|---|---|
| Ni | 6.59 |
| FeO | 12.60 |
| NiO | 10.03 |
| Fe | 7.07 |
| Change | − 2.09 |

It is interesting to investigate the order of magnitude of the pressure that would be required in order to produce an appreciable shift in the equilibrium. The volume decrease per mole of reactants is of the order of 2 cc. We then have, for the free-energy charge in increasing the pressure at constant temperature:[3]

[1] The thermodynamic nomenclature used in this discussion is that of Lewis and Randall (1923).

[2] A similar calculation based upon the densities of the silicates $2FeO \cdot SiO_2$ (Roth and Troitzsch, 1932–33, p. 79) and $2NiO \cdot SiO_2$ (Taylor, 1930, p. 241) gives a volume decrease of 2.27 cc. per mole of reactant metals.

[3] The symbol $K$ is used in this discussion to denote the true equilibrium constant. For the purposes of approximate calculation it is assumed that the mass-action constant, $C$, lies close to the true equilibrium constant.

$$\Delta F = RT\ln \frac{K_{\text{obs}}}{K_{p=0}} \qquad \begin{aligned} K_{\text{obs}} &= \text{Observed equilibrium constant for meteorites ;} \\ K_{p=0} &= \text{Equilibrium constant at zero pressure ;} \end{aligned}$$

$$\cong RT\ln 40 \ .$$

For a temperature of 2000° K. this corresponds to an energy change of 7.4 K cal/mole. Remembering that $\Delta F = \int \Delta V dP$ and that $\Delta V = 2$ cc/mole, we have

$$P \cong 1.5 \times 10^5 \text{ atm.}$$

Such a pressure is comparable to the internal pressures existing within Mars.

A more informative picture of the relationships between iron, nickel, and their silicates within meteorites results from a study of the iron and nickel contents of stony meteorites as a function of their metal-phase contents. The authors (1947*b*, pp. 508–510) have recently confirmed Prior's relationship concerning the nickel content of the metal phase by the process of arranging a series of ninety-five selected stony meteorite analyses in order of increasing metal-phase content and then averaging the nickel contents of the metal phases over specified intervals of metal-phase content. The results, reprinted from the authors' previous paper, are shown in figure 1. If now, instead of nickel concentrations, the average equilibrium constant within each range is plotted as a function of metal-phase concentration, the result shown in figure 2 is obtained.[4] From figure 2 it can be seen that as the metal-phase content increases, the equilibrium constant increases.

On the basis of the curve shown in figure 2, one must conclude that, if the iron and nickel concentrations in stony meteorites represent equilibrium concentrations, then *the conditions under which equilibrium was achieved varied from meteorite to meteorite in such a way that the greater the metal-phase content, the higher the temperature or the pressure.*

Additional information supporting the possibility that equilibrium at one time existed between the silicate and the metal phases can be obtained from a close inspection of the distribution of a series of elements between the two phases and by a comparison of the distribution coefficients with the heats of formation of the oxides (in the absence of adequate thermal data on the silicates). One would expect that if the heat of the reaction is the prime contributor to the free energy, then an approximation to a straight line would result when

$$\log \frac{(\text{Fe})_{\text{m}}\,(M)_{\text{si}}}{(M_{\text{m}})\ (\text{Fe})_{\text{si}}}$$

is plotted against the differences between the heats of formation of the metal oxides (per oxygen atom) and ferrous oxide.

Table 5 gives the data for the distribution of several elements between the silicate and metal phases, together with the heats of formation of their oxides. The results are plotted in figure 3. It can be seen that a straight line yielding a temperature of approximately 3000° C. is not an unreasonable approximation to the data, although the crudeness of such an estimation cannot be overemphasized. Here, as in the case of nickel alone, if pressure effects are, in reality, of considerable importance in establishing equilibrium, the shape of the curve shown in

[4] Unfortunately, there are very few analyses of the silicate phase of stony meteorites in which the NiO results can be trusted. However, a sufficient number of analyses has been accumulated to permit each point in fig. 2 to be represented by the average NiO content of three or more stony meteorites.

figure 3 and consequently the calculated temperature would be considerably altered.

In view of the strong indication that the equilibrium distribution of elements between the metal and silicate phases depends in large measure upon the relative heats of formation of the oxides (or silicates) involved, one might expect that the distribution of elements between the sulphide and metal phases would depend in like manner upon the relative heats of formation of the sulphides involved. Actually, close inspection of the meteoritic sulphide-metal distribution coefficients and thermal data reveals little or no correlation between the two quantities. However, if we compare the relative heats of formation of oxides and sulphides (table 6) and once again assume the existence of a pressure effect in establishing equilibrium, an explanation presents itself.

It can be seen from table 6 that the heats of formation of sulphides are considerably lower than the heats of formation of the corresponding oxides. As a result, the energy change in a reaction of the type

$$M + FeS \rightleftharpoons MS + Fe$$

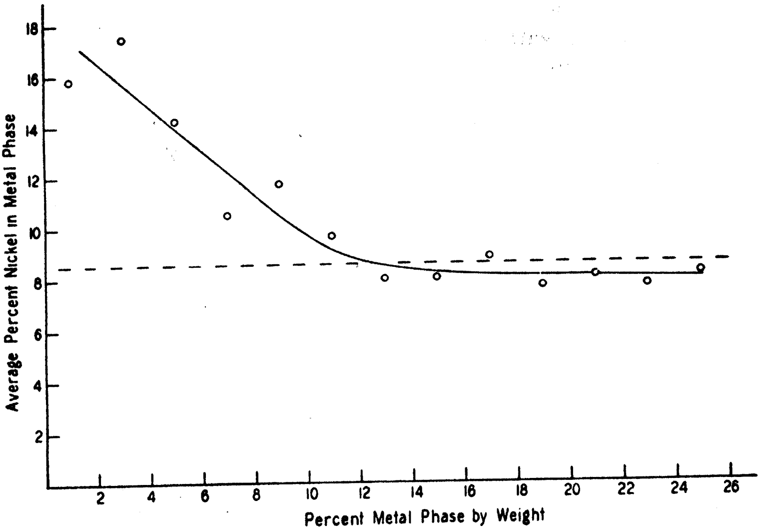



FIG. 1.—The variation of the nickel content in the metal phase of stony meteorites with the metal-phase content in the meteorite. The dotted line represents the average nickel content of iron meteorites. Each point represents the average of between two and nineteen stony meteorites.

will generally be considerably smaller than the energy change in a reaction of the type

$$M + FeSiO_3 \rightleftharpoons MSiO_3 + Fe\,.$$

Recalling the indications that a pressure-volume effect is required to explain the large equilibrium constant in the case of the iron-nickel-silicate equilibrium and remembering that in such a case $P\Delta V$ would be of the same order of magnitude as the chemical energy change, we see

that *in the case of the sulphides, where the heats of formation are small,* PΔV *would be the major factor in establishing equilibrium.*

Reference once again to the case of nickel can elucidate the point. Zur Strassen (1931) has measured equilibrium constants for the reaction.

$$Ni + FeS \rightleftharpoons NiS + Fe ,$$

and at about 1200° K. he found that the constant $C$ possesses a value very close to unity. If we now look at the observed distribution of nickel and iron between troilite and metal, we find that for meteorites

$$C \cong 0.02 ,$$

a value so low as to be practically unexplainable on a temperature-difference

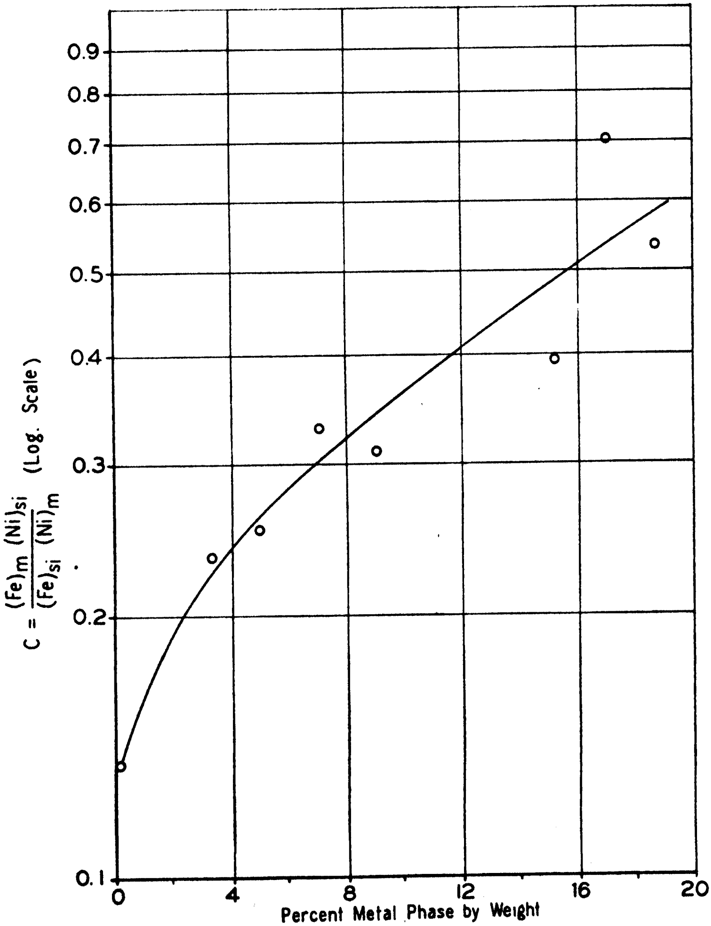



FIG. 2.—The iron-nickel equilibrium constant, $C = \frac{(Fe)_m(Ni)_{si}}{(Ni)_m(Fe)_{si}}$, in stony meteorites as a function of the metal-phase content.

basis. However, if the molar volumes of the components of the reaction are examined (table 7), it will be seen that a volume increase is associated with the reaction as it proceeds to the right. A pressure effect would, as a consequence, shift the equilibrium to the left.

definite proof must await refined thermochemical calculations, more adequate thermal data, more precise measurements of distribution coefficients, and a broader theoretical foundation of the effects of high pressures on partial molal volumes.

TABLE 5

HEATS OF FORMATION AND METEORITIC DISTRIBUTION COEFFICIENTS OF VARIOUS METAL OXIDES

| Element | Silicate Phase of Stony Meteorites | | Iron Meteorites | | $\frac{(Fe)_m(M)_{si}}{(Fe)_{si}(M)_m}$ | Compound | ΔH of Formation per Oxygen Atom (K Cal/Mole) | ΔH of Formation per Oxygen Atom Minus $(\Delta H)_{FeO}$ |
|---|---|---|---|---|---|---|---|---|
| | Parts per Million | Ref. | Parts per Million | Ref. | | | | |
| Ca | 20,700 | 1 | 500 | 3 | 280 | CaO | 152 | 86.3 |
| Mg | 166,000 | 1 | 320 | 3 | 3,500 | MgO | 144 | 78.3 |
| Al | 18,300 | 1 | 40 | 3 | 3,100 | $Al_2O_3$ | 133 | 67.3 |
| Ti | 1,000 | 1 | 100 | 4 | 68 | $TiO_2$ | 109 | 43.3 |
| V | 50 | 5 | 6.2 | 3 | 55 | $V_2O_3$ | 105 | 39.3 |
| Si | 216,000 | 1 | 8,000 | 3 | 180 | $SiO_2$ | 101 | 35.3 |
| Mn | 3,100 | 1 | 300 | 3 | 70 | MnO | 90.8 | 25.1 |
| Cr | 3,900 | 5 | 300 | 5 | 88 | $Cr_2O_3$ | 89.3 | 23.6 |
| Zr | 100 | 3 | 8 | 3 | 85 | $ZrO_2$ | 89.3 | 23.6 |
| Zn | 76 | 6 | 115 | 3 | 4.5 | ZnO | 85 | 19.3 |
| Sn | 4 | 3 | 102 | 3 | 0.27 | SnO | 69.8 | 4.1 |
| Fe | 132,000 | 1 | 897,930 | 2 | 1.0 | FeO | 65.7 | 0 |
| Mo | 2.5 | 3 | 16.6 | 3 | 1.0 | $MoO_2$ | 65.7 | 0 |
| Ni | 3,900 | 1 | 84,947 | 2 | 0.31 | NiO | 57.9 | −7.8 |
| Co | 200 | 1 | 6,230 | 2 | 0.22 | CoO | 57.5 | −8.2 |
| Pb | 2 | 5 | 53 | 3 | 0.26 | PbO | 52.5 | −13.2 |
| Cu | 1.6 | 3 | 305 | 3 | 0.036 | CuO | 35 | −30.7 |
| Pt | 0.08 | 3 | 19.4 | 6 | 0.029 | PtO | 17 | −48.7 |

REFERENCES FOR TABLE 5

1. Brown and Patterson, 1947*a*, pp. 405–411.
2. Brown and Patterson, 1947*b*, pp. 508–510.
3. Noddack, 1930, p. 757.
4. Ishibashi, 1931, p. 372.
5. Goldschmidt, 1938.
6. Noddack, 1934, p. 173.

It must be emphasized that any conclusions concerning the existence of pressure effects as main factors in determining meteoritic distribution coefficients are predicated on the assumption that the coefficients as observed are somewhat close to equilibrium values. Although existing data, such as that shown in figure 2, support such an assumption,

Nevertheless, if the validity of the assumption is admitted, the conclusions to be drawn on the basis of existing data appear straightforward:

1. Equilibrium was established under conditions of relatively high temperature and pressure (approximately 3000° C. and $10^5$–$10^6$ atm.).

2. The temperature and pressure ef-

fects were such that the greater the metal content of a fragment of meteoritic matter, the higher the temperature and/or the pressure at which equilibrium was attained.

In the light of these conclusions, the temptation becomes great to go further and to associate meteoritic matter with a planet in which the metal content, the temperature, and the pressure increased toward the center. This, of course, is much the same conclusion as that drawn so many years ago by Boisse; and, as will be shown below, the more closely one examines the data, the more clearly does such a conclusion appear to be justified.

### GENETIC RELATIONSHIPS

Of all major meteoritic constituents, the one which varies over the widest

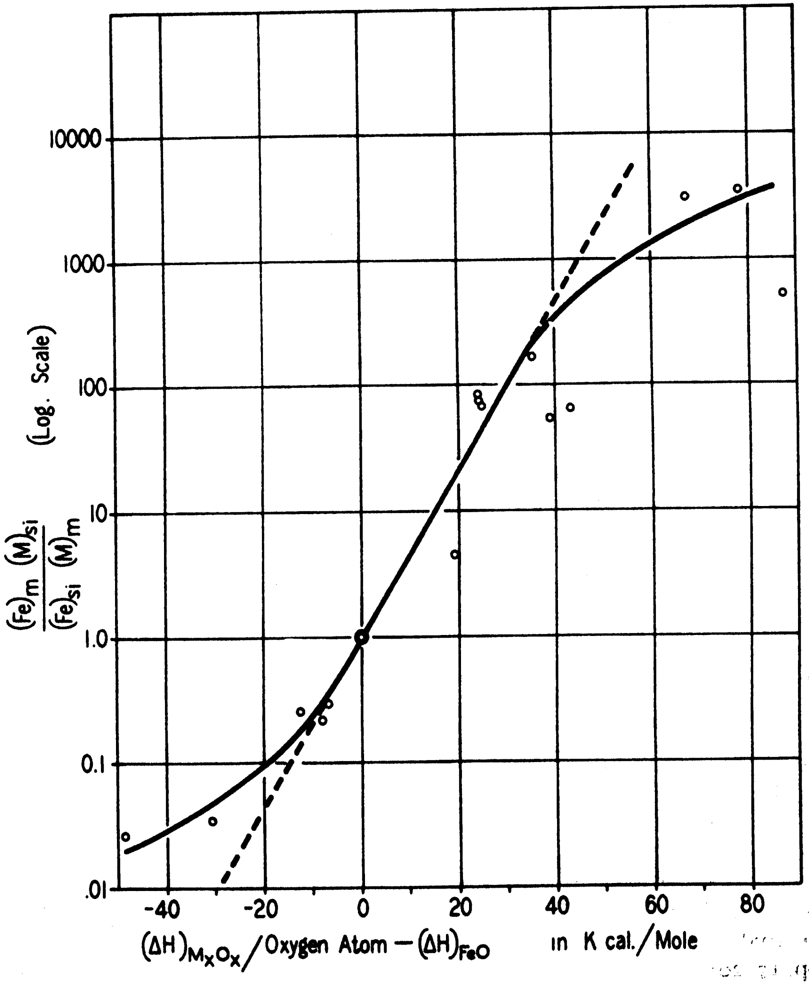



FIG. 3.—Observed meteoritic equilibrium constants, $\frac{(Fe)_m(M)_{si}}{(Fe)_{si}(M)_m}$, for various elements as a function of the heats of formation of their oxides.

range is metallic iron. Many meteorites are known in which the metal-phase content is close to zero; many are known in which the metal-phase content is 100 per cent; and specimens have been found representing every gradation between these two extremes. If, as our phase-equilibria studies indicate, meteorites actually came from a planet in which temperature, pressure, and metal-phase content increased toward the center, a study of gross meteoritic composition as a function of metal-phase content is of interest.

TABLE 6

HEATS OF FORMATION OF A FEW METAL OXIDES AND SULPHIDES

| Compound | Heat of Formation (K Cal/Mole) | Compound | Heat of Formation (K Cal/Mole) |
|---|---|---|---|
| FeO | 65.7 | FeS | 23.1 |
| NiO | 57.9 | NiS | 17.4 |
| CoO | 57.5 | CoS | 19.7 |
| CuO | 34.9 | CuS | 11.6 |
| ZnO | 85.0 | ZnS | 45.9 |
| SnO | 69.8 | SnS | 23.9 |
| CdO | 65.2 | CdS | 33.9 |
| PbO | 52.5 | PbS | 22.2 |

A series of 127 selected stony-meteorite analyses was arranged in order of increasing metal-phase content, and the gross composition was then averaged over specific intervals of metal-phase content, thus obtaining the data shown in table 8. In figures 4, 5, and 6 the data are presented on a weight percentage basis, and in figures 7 and 8 some of the data are presented on an atomic percentage basis. For convenience, corresponding data for the average composition of igneous rocks (Clarke and Washington, 1922, p. 108) and plateau basalt (Daly, 1933, pp. 17, 201) are presented, together with the meteoritic data.

From the figures several interesting features become apparent. As the metal-phase content increases:

1. Combined iron[5] increases, passing through a maximum at a metal-phase content of 3–4 per cent, following which the combined iron decreases steadily.

2. Total iron (and total iron, nickel, and cobalt) increases rapidly until a metal-phase content of 3–4 per cent is reached, following which the increase with metal-phase content is more gradual.

3. Oxygen decreases steadily.

4. Silicon decreases rapidly until a metal-phase content of 3–4 per cent is reached, following which the silicon concentration levels off.

5. Magnesium increases, leveling off at a metal-phase content of approximately 5 per cent.

6. Calcium decreases, becoming constant at a metal-phase content of approximately 5 per cent.

7. Aluminum decreases, leveling off at a metal-phase content of approximately 5 per cent.

8. Sulphur increases rapidly until a metal-phase content of 3–4 per cent is reached, following which the general tendency is for the sulphur to remain constant.

TABLE 7

VOLUME CHANGE IN THE REACTION $Ni + FeS \rightleftarrows NiS + Fe$

| Substance | Molar Volume (cc.) |
|---|---|
| Ni | 6.59 |
| FeS | 18.16 |
| NiS | 19.73 |
| Fe | 7.07 |
| Change | + 2.05 |

One of the more interesting features of the data is the relationship between igneous rock, plateau basalt, and stony

[5] "Combined iron" is iron combined as oxides and sulphides.

## TABLE 8

AVERAGE COMPOSITION OF STONY METEORITES AS A FUNCTION OF METAL-PHASE CONTENT
("Wt. %" Denotes Per Cent by Weight. "A%" Denotes Atomic Per Cent)

| Metal-Phase Interval Per Cent by Weight | | Combined Iron | Metallic Iron | Total Iron | Combined Nickel | Total Nickel | Total Cobalt | Total Iron, Nickel, Cobalt | Oxygen | Silicon | Magnesium | Aluminum | Calcium | Sulphur | Sodium | Potassium | Titanium | Phosphorus | Manganese | Chromium | Metal Phase | Number of Analyses |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Igneous rock... | Wt.% | 5.10 | ...... | 5.10 | 0.01 | 0.01 | ..... | 5.11 | 46.49 | 27.62 | 2.11 | 8.12 | 3.63 | 0.05 | 2.85 | 2.60 | 0.63 | 0.13 | 0.09 | 0.04 | ...... | .... |
| | A% | 1.94 | ...... | 1.94 | ..... | ..... | ..... | 1.94 | 61.83 | 20.95 | 1.85 | 6.41 | 1.93 | ..... | 2.64 | 1.42 | 0.28 | 0.09 | 0.03 | ..... | ...... | |
| Plateau basalt. | Wt.% | 8.09 | ...... | 8.09 | ..... | ..... | ..... | 8.09 | 45.41 | 23.64 | 2.95 | 9.21 | 5.78 | ..... | 2.40 | 1.46 | 0.41 | 0.09 | 0.36 | ..... | ...... | .... |
| | A% | 3.10 | ...... | 3.10 | ..... | ..... | ..... | 3.10 | 60.78 | 18.04 | 2.60 | 7.31 | 3.09 | ..... | 2.24 | 0.80 | 0.18 | 0.06 | 0.14 | ..... | ...... | |
| 0– 1... | Wt.% | 14.04 | 0.19 | 14.23 | 0.37 | 0.41 | 0.04 | 14.69 | 41.58 | 21.92 | 10.93 | 4.31 | 6.08 | 0.61 | 0.51 | 0.12 | 0.40 | 0.07 | 0.27 | 0.28 | 0.23 | 14 |
| | A% | 5.54 | 0.08 | 5.61 | ..... | 0.15 | 0.02 | 5.78 | 57.26 | 17.21 | 9.90 | 3.52 | 3.34 | 0.42 | 0.49 | 0.07 | 0.18 | 0.05 | 0.11 | 0.12 | ...... | |
| 1– 2... | Wt.% | 14.79 | 1.06 | 15.85 | ..... | 0.21 | 0.02 | 16.08 | 41.39 | 21.87 | 12.22 | 2.98 | 3.29 | 1.01 | 0.47 | 0.19 | 0.55 | 0.05 | 0.23 | 0.28 | 1.29 | 3 |
| | A% | 5.97 | 0.43 | 6.41 | ..... | 0.08 | 0.01 | 6.50 | 58.41 | 17.60 | 11.35 | 2.50 | 1.85 | 0.71 | 0.46 | 0.11 | 0.26 | 0.04 | 0.09 | 0.12 | ...... | |
| 2– 3... | Wt.% | 17.63 | 1.87 | 19.50 | 0.65 | 1.05 | 0.21 | 20.66 | 38.70 | 19.10 | 16.04 | 1.13 | 1.26 | 1.75 | 0.63 | 0.08 | ..... | 0.10 | 0.22 | 0.25 | 2.30 | 5 |
| | A% | 7.34 | 0.78 | 8.12 | ..... | 0.42 | 0.08 | 8.62 | 56.26 | 15.83 | 15.34 | 0.97 | 0.73 | 1.27 | 0.64 | 0.05 | ..... | 0.07 | 0.09 | 0.11 | ...... | |
| 3– 4... | Wt.% | 20.67 | 2.78 | 23.45 | 0.84 | 1.44 | 0.10 | 24.99 | 37.78 | 18.00 | 11.86 | 1.34 | 2.16 | 2.14 | 0.80 | 0.18 | ..... | 0.29 | 0.42 | 0.42 | 3.43 | 4 |
| | A% | 8.87 | 1.19 | 10.07 | ..... | 0.59 | 0.04 | 10.69 | 56.60 | 15.38 | 11.69 | 1.19 | 1.29 | 1.60 | 0.83 | 0.11 | ..... | 0.23 | 0.18 | 0.19 | ...... | |
| 4– 6... | Wt.% | 16.13 | 4.28 | 20.41 | 0.55 | 1.27 | 0.13 | 21.81 | 38.76 | 17.70 | 15.24 | 2.21 | 1.39 | 1.73 | 0.68 | 0.20 | 0.10 | 0.10 | 0.16 | 0.31 | 5.05 | 14 |
| | A% | 6.73 | 1.79 | 8.52 | ..... | 0.50 | 0.05 | 9.08 | 56.49 | 14.71 | 14.61 | 1.91 | 0.81 | 1.26 | 0.69 | 0.12 | 0.05 | 0.07 | 0.07 | 0.14 | ...... | |
| 6– 8... | Wt.% | 16.43 | 6.33 | 22.76 | 0.51 | 1.26 | 0.08 | 24.10 | 38.03 | 18.37 | 14.86 | 1.29 | 1.16 | 1.88 | 0.69 | 0.14 | 0.11 | 0.17 | 0.17 | 0.27 | 7.13 | 11 |
| | A% | 6.88 | 2.65 | 9.53 | ..... | 0.50 | 0.03 | 10.07 | 55.99 | 15.31 | 14.29 | 1.12 | 0.68 | 1.37 | 0.70 | 0.08 | 0.05 | 0.13 | 0.07 | 0.12 | ...... | |

TABLE 8—*Continued*

| Metal-Phase Interval Per Cent by Weight | | Combined Iron | Metallic Iron | Total Iron | Combined Nickel | Total Nickel | Total Cobalt | Total Iron, Nickel, Cobalt | Oxygen | Silicon | Magnesium | Aluminum | Calcium | Sulphur | Sodium | Potassium | Titanium | Phosphorus | Manganese | Chromium | Metal Phase | Number of Analyses |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 8–10... | Wt.% | 15.06 | 7.96 | 23.02 | 0.46 | 1.53 | 0.11 | 24.66 | 36.27 | 18.40 | 14.24 | 1.28 | 1.40 | 2.26 | 0.73 | 0.20 | 0.09 | 0.14 | 0.40 | 0.33 | 9.09 | 19 |
| | A% | 6.48 | 3.43 | 9.91 | ..... | 0.63 | 0.05 | 10.59 | 54.52 | 15.77 | 14.08 | 1.14 | 0.84 | 1.70 | 0.76 | 0.12 | 0.05 | 0.11 | 0.18 | 0.15 | ...... | |
| 10–12... | Wt.% | 14.16 | 9.97 | 24.13 | 0.15 | 1.22 | 0.08 | 25.43 | 36.44 | 18.36 | 14.38 | 1.32 | 1.23 | 1.83 | 0.56 | 0.33 | 0.04 | 0.16 | 0.29 | 0.26 | 11.10 | 8 |
| | A% | 6.09 | 4.29 | 10.38 | ..... | 0.50 | 0.03 | 10.91 | 54.71 | 15.72 | 14.20 | 1.17 | 0.74 | 1.37 | 0.58 | 0.20 | 0.02 | 0.12 | 0.13 | 0.12 | ...... | |
| 12–14... | Wt.% | 13.29 | 11.90 | 25.19 | 0.24 | 1.28 | 0.09 | 26.56 | 35.68 | 17.92 | 14.32 | 1.62 | 1.28 | 1.57 | 0.97 | 0.13 | 0.02 | 0.07 | 0.20 | 0.35 | 13.03 | 6 |
| | A% | 5.76 | 5.16 | 10.92 | ..... | 0.53 | 0.04 | 11.48 | 53.98 | 15.46 | 14.25 | 1.45 | 0.77 | 1.19 | 1.02 | 0.08 | 0.01 | 0.06 | 0.09 | 0.16 | ...... | |
| 14–16... | Wt.% | 13.36 | 13.95 | 27.31 | 0.33 | 1.56 | 0.10 | 28.97 | 33.73 | 17.59 | 13.23 | 1.88 | 1.46 | 2.19 | 0.56 | 0.35 | 0.02 | 0.19 | 0.58 | 0.26 | 15.24 | 6 |
| | A% | 5.94 | 6.20 | 12.15 | ..... | 0.66 | 0.04 | 12.85 | 52.36 | 15.57 | 13.51 | 1.73 | 0.90 | 1.70 | 0.61 | 0.22 | 0.01 | 0.15 | 0.26 | 0.12 | ...... | |
| 16–18... | Wt.% | 11.77 | 15.49 | 27.26 | 0.56 | 2.05 | 0.13 | 29.44 | 33.47 | 16.85 | 13.74 | 1.51 | 1.39 | 2.02 | 1.04 | 0.23 | 0.06 | 0.12 | 0.76 | 0.32 | 17.06 | 10 |
| | A% | 5.25 | 6.91 | 12.16 | ..... | 0.87 | 0.05 | 13.09 | 52.14 | 14.97 | 14.08 | 1.40 | 0.86 | 1.57 | 1.13 | 0.15 | 0.03 | 0.10 | 0.34 | 0.15 | ...... | |
| 18–20... | Wt.% | 10.46 | 17.18 | 27.64 | 0.32 | 1.75 | 0.11 | 29.50 | 33.51 | 17.12 | 14.34 | 1.33 | 1.29 | 1.85 | 0.73 | 0.12 | 0.12 | 0.13 | 0.20 | 0.20 | 18.72 | 13 |
| | A% | 4.67 | 7.67 | 12.35 | ..... | 0.74 | 0.05 | 13.14 | 52.25 | 15.22 | 14.71 | 1.23 | 0.80 | 1.44 | 0.79 | 0.08 | 0.06 | 0.10 | 0.09 | 0.09 | ...... | |
| 20–22... | Wt.% | 9.45 | 18.93 | 28.38 | 0.08 | 1.74 | 0.09 | 30.21 | 32.80 | 17.10 | 14.07 | 1.32 | 1.46 | 2.03 | 0.59 | 0.23 | 0.10 | 0.11 | 0.19 | 0.22 | 20.67 | 7 |
| | A% | 4.26 | 8.54 | 12.80 | ..... | 0.75 | 0.04 | 13.58 | 51.63 | 15.35 | 14.57 | 1.23 | 0.92 | 1.59 | 0.65 | 0.15 | 0.05 | 0.09 | 0.09 | 0.11 | ...... | |
| 22–24... | Wt.% | 9.14 | 21.16 | 30.30 | 0.16 | 1.94 | 0.12 | 32.36 | 31.52 | 16.08 | 13.97 | 1.51 | 1.31 | 2.14 | 0.69 | 0.12 | 0.05 | 0.08 | 0.26 | 0.18 | 23.06 | 4 |
| | A% | 4.20 | 9.73 | 13.93 | | 0.85 | 0.05 | 14.83 | 50.57 | 14.71 | 14.74 | 1.44 | 0.84 | 1.71 | 0.77 | 0.08 | 0.03 | 0.07 | 0.12 | 0.09 | | |
| 24–26... | Wt.% | 8.90 | 22.48 | 31.38 | 0.09 | 2.09 | 0.13 | 33.60 | 30.71 | 15.90 | 13.71 | 1.24 | 1.37 | 2.48 | 0.78 | 0.16 | 0.05 | 0.03 | 0.30 | 0.18 | 24.61 | 3 |
| | A% | 4.13 | 10.44 | 14.57 | ..... | 0.92 | 0.06 | 15.55 | 49.78 | 14.69 | 14.62 | 1.19 | 0.89 | 2.01 | 0.88 | 0.11 | 0.03 | 0.03 | 0.14 | 0.09 | ...... | |

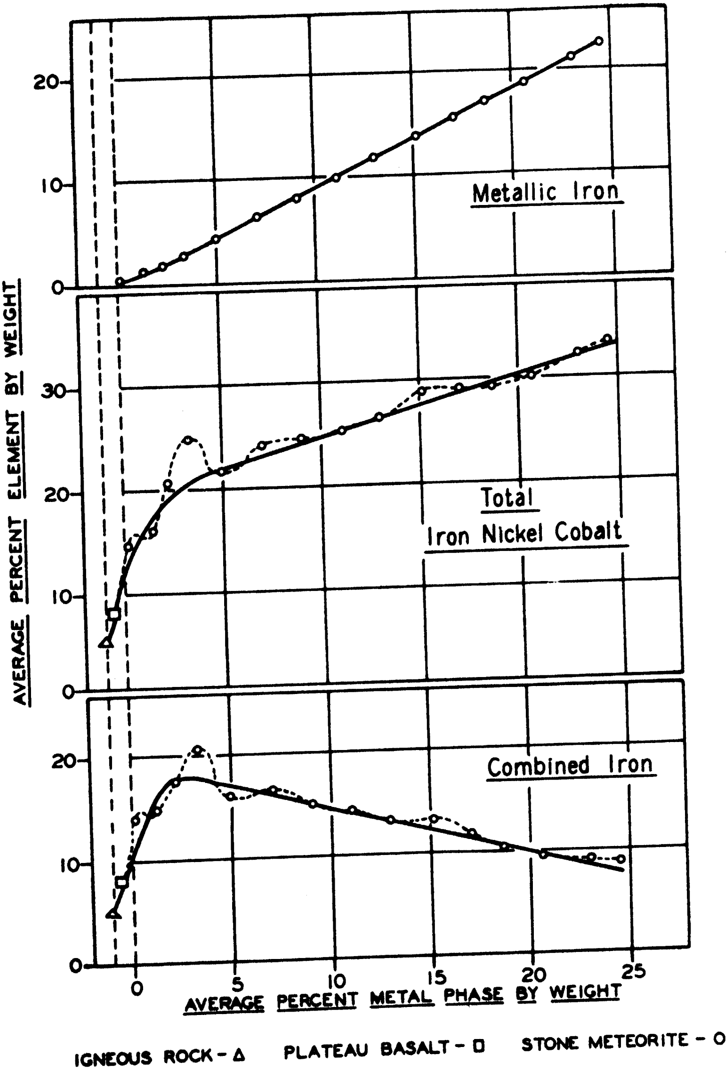



FIG. 4.—Variation of the weight percentages of metallic iron, total iron, nickel, and cobalt and combined iron in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparison.

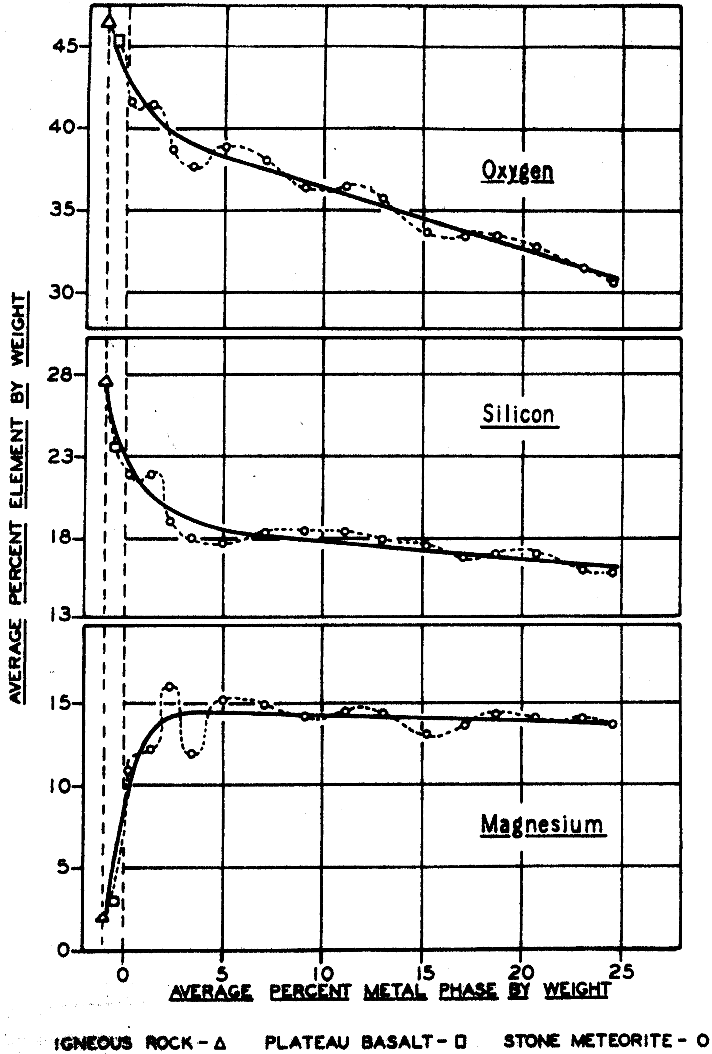



FIG. 5.—Variation of the weight percentages of oxygen, silicon, and magnesium in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparison.

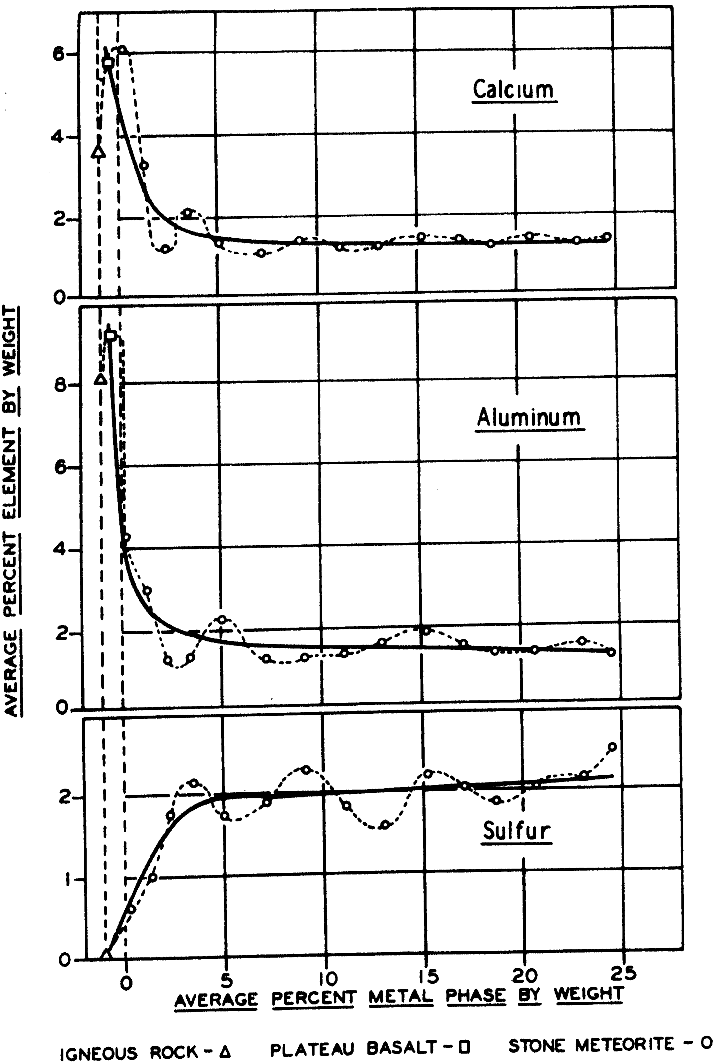



FIG. 6.—Variation of the weight percentages of calcium, aluminum, and sulphur in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparison.

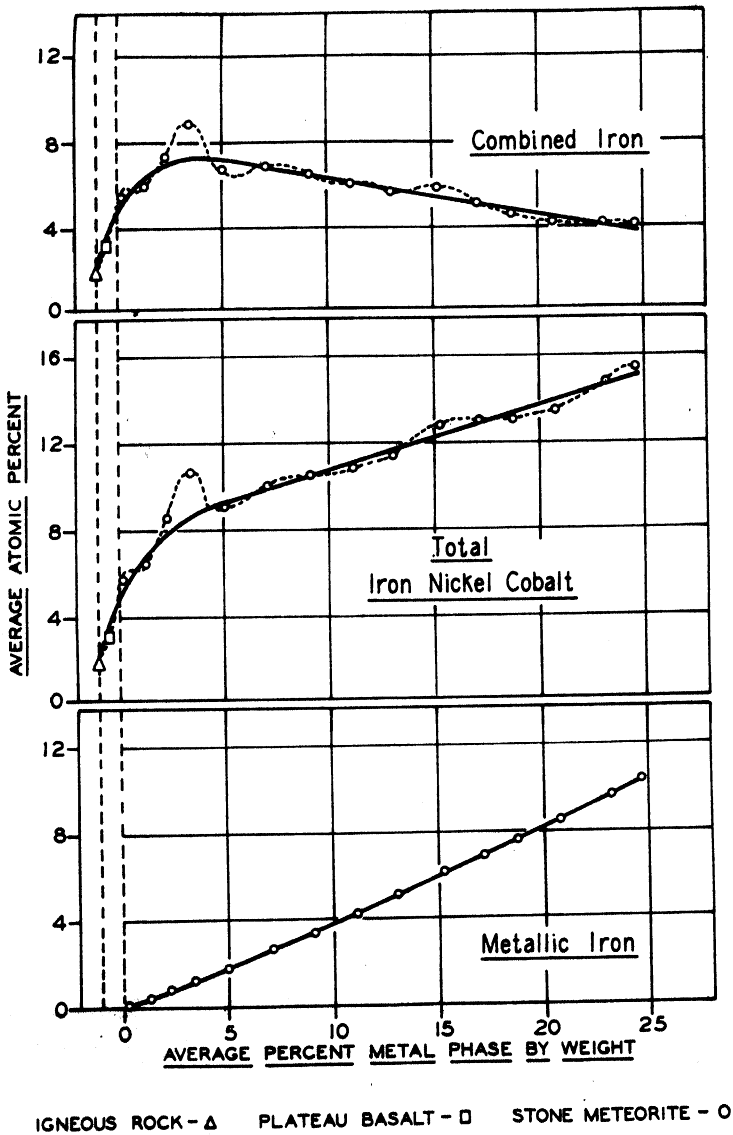



FIG. 7.—Variation of the atomic percentages of combined iron, total iron, nickel, and cobalt and metallic iron in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparison.

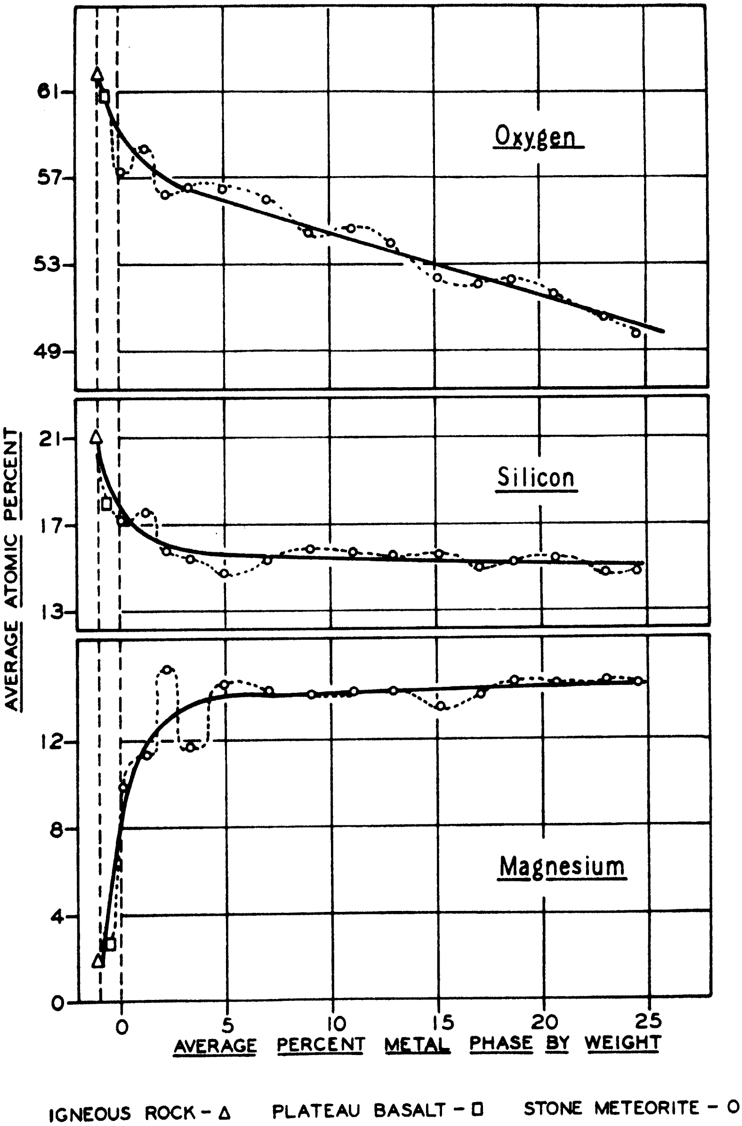



FIG. 8.—Variation of the atomic percentages of oxygen, silicon, and magnesium in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparison.

meteorites which possess low metal-phase contents. *In the cases of combined iron, total iron, oxygen, silicon, magnesium, and calcium, the plateau basalt values are clearly intermediate between the corresponding values for igneous rock and meteorites.* Aluminum is a definite exception to this regularity; and in the case of sulphur the lack of data for plateau basalt prevents comparison.

TABLE 9

RELATIVE ABUNDANCE OF TRANSITION METALS IN STONY METEORITES

| Metal-Phase Range (Weight Per Cent) | Av. Total Iron, Nickel, Cobalt (Atomic Per Cent) | Av. Oxygen (Atomic Per Cent) | Sum of Avs. |
|---|---|---|---|
| 0– 1 | 5.78 | 57.26 | 63.04 |
| 1– 2 | 6.50 | 58.41 | 64.91 |
| 2– 3 | 8.62 | 56.20 | 64.88 |
| 3– 4 | 10.69 | 56.60 | 67.29 |
| 4– 6 | 9.08 | 56.49 | 65.57 |
| 6– 8 | 10.07 | 55.99 | 66.06 |
| 8–10 | 10.59 | 54.52 | 65.11 |
| 10–12 | 10.91 | 54.71 | 65.62 |
| 12–14 | 11.48 | 53.98 | 65.46 |
| 14–16 | 12.85 | 52.36 | 65.21 |
| 16–18 | 13.09 | 52.14 | 65.23 |
| 18–20 | 13.14 | 52.25 | 65.39 |
| 20–22 | 13.58 | 51.63 | 65.21 |
| 22–24 | 14.83 | 50.57 | 65.40 |
| 24–26 | 15.55 | 49.78 | 65.33 |

Indeed, it seems improbable that such a marked relationship between igneous rock, plateau basalt, and meteorites is fortuitous. Again one is forced to the conclusion that meteorites once formed a planet similar in general physicochemical characteristics to the earth.

Inspection of figures 4–8 serves to show that at metal-phase contents greater than 4–5 per cent, the main variables in meteorites are iron and oxygen, with the decrease in oxygen almost exactly matching the increase in iron, cobalt, and nickel, on an atomic percentage basis. This relationship between total iron, cobalt, and nickel and oxygen is of remarkable constancy, as can be seen from table 9, where the sums of the atomic percentages of these metals and oxygen are tabulated. It seems clear that at metal-phase contents greater than 4–5 per cent, the gradation from one meteorite to another consists primarily of the removal of atoms of oxygen (from atoms of combined iron, cobalt, and nickel) and the replacement of the oxygen atoms with equal numbers of atoms of metallic iron, cobalt, and nickel. Of course, as a result of this relationship between oxygen and the transition elements, the ratio of combined iron to metallic iron increases rapidly as the metal-phase content decreases. This latter effect is illustrated in figure 9.

Silicon and magnesium are rather constant at metal-phase contents greater than 5 per cent. Perhaps the most interesting feature of these two elements is their atomic ratios, which lie close to unity over a wide range of metal-phase concentrations. This is illustrated in table 10.

Concerning the interplay of one element with another, very little can be said in addition to the major effects already discussed. A few points should be mentioned, however, in connection with the prominent oscillations noted in figures 4–8, best exemplified by the oxygen and sulphur curves (figs. 5 and 6). It can be seen quite clearly that when oxygen is a maximum, sulphur is a minimum. A similar effect appears in the case of magnesium (fig. 5), where it will be noted that magnesium maxima and minima coincide with those of oxygen. Combined iron and calcium, on the other hand, appear to possess curves similar to those of sulphur, where the minima coincide with the oxygen maxima. Neither the silicon nor the

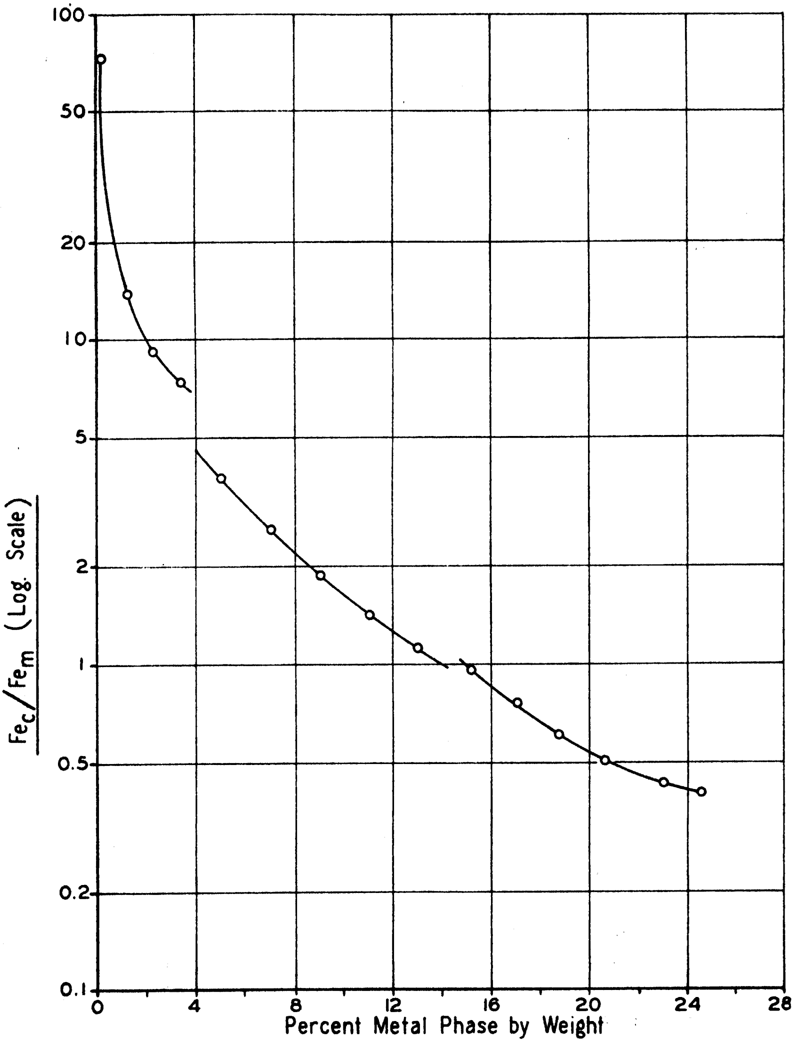



FIG. 9.—The ratio of combined to uncombined iron as a function of metal-phase concentration

aluminum curves appear to bear any relationship (with respect to maxima and minima) to the other curves, although there is some indication of an interplay between these two elements.

Summarizing these minor effects, the composition curves of the elements denoted in figures 4–8 can be divided into three classes:

| Class 1 | Class 2 | Class 3 |
|---|---|---|
| Oxygen | Sulphur | Silicon |
| Magnesium | Combined iron | Aluminum |
| | Calcium | |

The curves within each class appear to possess similar structures, and the maxima of class 1 coincide with the minima of class 2. Class 3 appears to be unrelated to either of the other two classes.

## INTERPRETATION

Perhaps the most significant feature of the relationships outlined in the preceding section is the observed phenomenon of the rapid increase in the oxidation state of meteorites as their metal-phase contents and total iron contents decrease. It is important that this effect be explained in more concrete terms than the explanation of Prior to the effect that meteorites have separated from a single magma which has passed through successive stages of progressive oxidation. We must inquire into the conditions that might have brought about such marked differences in oxidation states.

Let us suppose, first, that meteorites are fragments of what was once a planet and that the planet was at one time in a liquid state. Let us suppose, further, that all the oxygen associated with this hypothetical planet was combined with the major metallic constituents, notably silicon, magnesium, and iron. Let us further suppose that metallic iron, cobalt, and nickel existed in excess, over and above the amount of oxygen available for combination. Under such circumstances one would expect the liquid to separate into two phases: metallic nickel-iron and magnesium and iron silicates. One would expect to find, in addition, a certain amount of metallic nickel-iron dissolved in the silicate phase. If the planet was small, one would expect the silicate phase to possess a uniform composition throughout and, neglecting surface crystallization effects, the ratio of combined iron to metallic iron to be constant from point to point. If we assume, on the other hand, that the planet was sufficiently large to possess significant gravitational effects, it is reasonable to suppose that a gradient in the dissolved metal would be set up in such a way that the metal content would increase as one approached the center. Unfortunately, however, such a simple picture cannot explain the rapid decrease in combined iron concentration with increasing metal-phase concentration. We must ask whether there is any basis for expecting the combined iron

TABLE 10

THE ATOMIC RATIO OF SILICON TO MAGNESIUM AS A FUNCTION OF METAL-PHASE CONTENT

| Metal-Phase Range (Per Cent by Wt.) | Atomic Ratio |
|---|---|
| 0– 1 | 1.74 |
| 1– 2 | 1.55 |
| 2– 3 | 1.03 |
| 3– 4 | 1.32 |
| 4– 6 | 1.01 |
| 6– 8 | 1.07 |
| 8–10 | 1.12 |
| 10–12 | 1.11 |
| 12–14 | 1.08 |
| 14–16 | 1.15 |
| 16–18 | 1.06 |
| 18–20 | 1.03 |
| 20–22 | 1.05 |
| 22–24 | 0.99 |
| 24–26 | 1.00 |

concentration to decrease with increasing depth within a planet.

It was demonstrated in the first part of this paper that pressures of the order of $10^5$–$10^6$ atm. can give rise to significant effects in certain equilibria, notably nickel. Similarly, pressure effects might be expected to affect the ratio of combined iron to metallic iron within a planet, if we take into account the large volume occupied by oxygen ions within the silicate phase and the relatively small density of those ions. For comparison table 11 gives the atomic and ionic radii of several elements encountered in meteorites, and table 12 shows the relative volumes occupied by the various elements within the silicate phase. The large relative volume occupied by oxygen is apparent.

In view of the very small density of oxygen ions, the effect of a large pressure gradient would be to "squeeze" the oxygen toward the outside of the planet. That such a squeezing process could actually break chemical bonds can be seen from the relationship for the chemical potential per gram of a given constituent present in the magma,

$$\frac{d\mu}{M} = (v_x - v_m)\, dP\,,$$

TABLE 11

IONIC AND ATOMIC RADII OF SOME METEORITIC ELEMENTS

| Element | Radius A° | Ref. | Element | Radius A° | Ref. |
|---|---|---|---|---|---|
| Fe | 1.24 | 3 | $Al^{+++}$ | 0.57 | 1 |
| $Fe^{++}$ | 0.75 | 2 | $Ca^{++}$ | 1.06 | 1 |
| Ni | 1.24 | 3 | $S^{--}$ | 1.74 | 1 |
| $Ni^{++}$ | 0.70 | 2 | $Na^{+}$ | 0.98 | 1 |
| Co | 1.25 | 3 | $K^{+}$ | 1.33 | 1 |
| $Co^{++}$ | 0.72 | 2 | $Ti^{++++}$ | 0.64 | 1 |
| $O^{--}$ | 1.40 | 2 | $P^{+++++}$ | 0.34 | 2 |
| $Si^{++++}$ | 0.40 | 1 | $Mn^{++}$ | 0.80 | 2 |
| $Mg^{++}$ | 0.78 | 1 | $Cr^{+++}$ | 0.64 | 2 |

REFERENCES FOR TABLE 11

1. Goldschmidt, 1927, p. 1263.
2. Pauling, 1940, p. 350 (Pauling's correction of Goldschmidt's data).
3. Neuburger, 1936, p. 1.

TABLE 12

RELATIVE VOLUMES OCCUPIED BY ELEMENTS IN METEORITES

| Metal-Phase Interval (Per Cent by Weight) | Total Iron | Total Nickel | Total Cobalt | Total Iron, Nickel, Cobalt | Oxygen | Silicon | Magnesium | Aluminum | Calcium | Sulphur | Sodium | Potassium | Titanium | Phosphorous | Manganese | Chromium |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Igneous rock | 0.46 | .... | .... | 0.46 | 94.19 | 0.75 | 0.49 | 0.66 | 1.28 | 0.15 | 1.38 | 1.86 | 0.04 | .... | 0.01 | 0.00 |
| 0–1 | 1.44 | 0.17 | 0.02 | 1.63 | 90.64 | 0.64 | 2.72 | 0.38 | 2.29 | 1.28 | 0.27 | 0.09 | 0.03 | .... | 0.03 | 0.02 |
| 8–10 | 5.19 | 0.67 | 0.06 | 5.91 | 83.47 | 0.56 | 3.74 | 0.12 | 0.56 | 5.01 | 0.40 | 0.16 | 0.01 | .... | 0.05 | 0.02 |
| 24–26 | 11.98 | 0.97 | 0.07 | 13.02 | 75.40 | 0.52 | 3.84 | 0.12 | 0.59 | 5.85 | 0.46 | 0.14 | 0.01 | .... | 0.04 | 0.01 |

where $v_x$ is the specific volume of a given constituent (as, for example, oxygen ions), $v_m$ is the specific volume of the magma, and $P$ is the pressure. Within a sphere large enough to give rise to internal pressures of the order of $10^5$–$10^6$ atm., the pressure gradients would be ample to produce shifts in the metallic iron–combined iron equilibrium. The specific volumes of oxygen ions, ferrous ions, and metallic iron atoms are, respectively, 0.4, 0.02, and 0.08 cc/gm. These can be compared with the specific volume of the magma, which is approximately 0.2 cc/gm.

Thus, if iron and iron oxide are to coexist at high pressures and in the presence of a pressure gradient, then necessarily at equilibrium the ratio of combined iron to metallic iron must decrease with increasing pressure. On the basis of such a picture, an increasing combined iron concentration associated with a decreasing metal-phase concentration would be expected. In addition, of course, when the ratio of combined iron to metallic iron becomes so large that the bulk of the iron is in the combined state, the combined iron concentration should pass through a maximum, following which it should decrease with decreasing metal concentration.[6]

Any interpretation of the features of the abundance distribution curves of the other elements encountered in meteorites (silicon, magnesium, calcium, etc.) must necessarily depend upon the detailed chemistry of silicates as a function of temperature and of pressure. However, a few broad features of the curves are reasonably apparent and can be interpreted on the basis of existing data.

The fact that the atomic ratio of silicon to magnesium lies so close to unity for the greater part of the metal-phase range is perhaps significant. The stabilities of magnesium silicates are well known. However, the absolute abundance of magnesium is somewhat less than that of silicon; and, as a result, there is not enough magnesium to combine with silicon on a one-to-one basis throughout. There results an excess of $SiO_2$, which, being less dense than the magnesium silicates, is squeezed to the outer portions of the planet, together with the silicates of calcium, aluminum, sodium, and potassium.

Crystallization effects, of course, greatly complicate the situation at low metal-phase concentrations. It is noteworthy, however, that magnesium apparently exists on the surface of the earth only through the courtesy of statistical fluctuations. As can be seen from figure 10, the frequency-distribution curve for magnesium in igneous rocks (Richardson and Sneesby, 1922, p. 303) is of a markedly different type from the corresponding distribution curve for magnesium in the silicate phase of stony meteorites. While the curve for meteorites centers around a well-defined maximum, the curve for igneous rock is exponential in character, with the majority of rocks possessing little magnesium. In other words, magnesium has in some manner been excluded either before or during the process of rock crystallization.

A similar correlation holds for other elements common to both igneous rocks and meteorites. In general, if the igneous-rock value for a given element is lower than the value for meteorites of low metal-phase content, the igneous-rock frequency-distribution curve will be exponential in character, with the highest value being at zero per cent. Conversely, if the igneous-rock value is higher than

[6] T. F. W. Barth has discussed the variation of oxygen content with depth in the earth's lithosphere (1948, p. 43).

the value for meteorites of low metal-phase content, the igneous rock frequency-distribution curve will be bell-shaped. The first case is exemplified by magnesium, calcium, and combined iron and the second case by silicon, aluminum, and sodium.

formation is coupled with the chemical relationships existing between igneous rock, plateau basalt, and stony meteorites, the simplest conclusion to be drawn is that meteorites are fragments of a planet similar to the earth in general physicochemical characteristics.

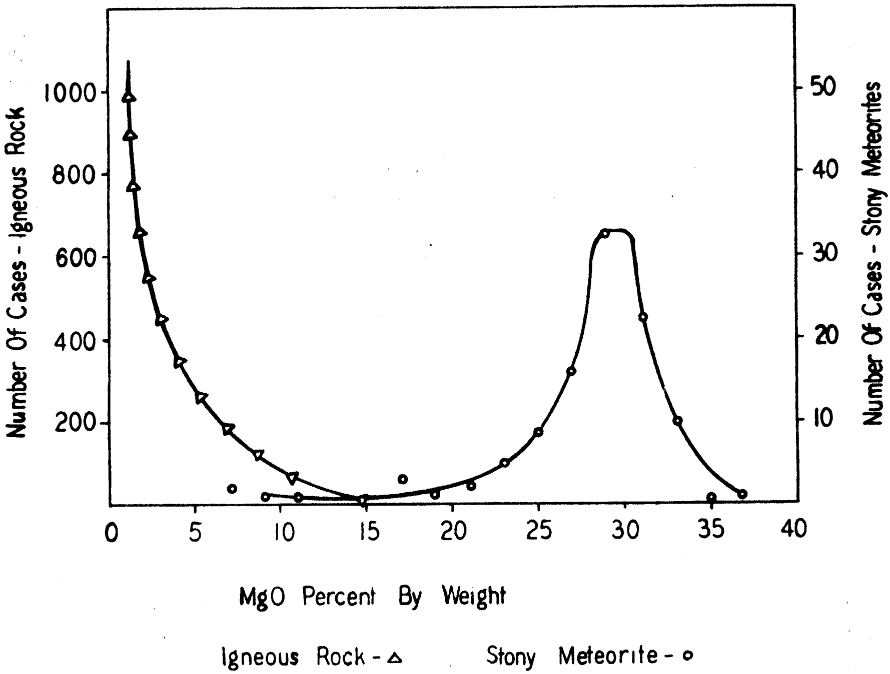



FIG. 10.—Frequency-distribution curves for magnesium oxide in igneous rocks and in the silicate phase of stony meteorites.

### PLANET STRUCTURE

We have seen thus far that there exist certain chemical regularities among meteorites involving oxidation-reduction equilibria (or at least something approaching equilibria). It has been demonstrated, furthermore, that the gradations in equilibria from meteorite to meteorite are difficult to explain without assuming the existence of pressure effects upon the equilibria, the order of magnitude of the pressure effects being most probably in the range of $10^5$–$10^6$ atm. When this information is coupled with the chemical relationships existing between igneous rock, plateau basalt, and stony meteorites, the simplest conclusion to be drawn is that meteorites are fragments of a planet similar to the earth in general physicochemical characteristics.

It is possible mentally to construct systems other than a planet which will explain some of the relationships encountered in the sections on phase equilibria in meteorites, and on genetic relationships. However, no system of which the authors have conceived has explained the data nearly so well as the assumption of a body of planetary dimensions. The existing data admittedly are sketchy; and perhaps more intensive study of meteorites in the laboratory will lead to contradictions of the data used in this discussion.

At that time the conclusions drawn here might be changed. However, until that time and to the extent that existing meteoritic data can be trusted, the conclusion appears irrefutable that meteorities at one time were an integral part of a planet. If the conclusion is correct, it is only proper to ask, "What were the properties of the planet?" In order to answer the question, we must investigate the relative frequency of meteorite falls as a function of composition.

The difference in behavior between iron and stony meteorites as they pass through the atmosphere and the differences due to weathering and ease of recognition are well known (Watson, 1939, p. 426). Because of these differences it is most difficult to assess, with any degree of precision, the absolute ratio of iron to stony meteorites striking the earth. However, within the class of stony meteorites it is only reasonable to suppose that such marked differences in behavior do not exist and that a study of frequency as a function of composition might have some significance.

In figure 11 a total of 184 selected meteorite metal-phase analyses[7] have been broken down into numbers of cases versus metal-phase content. Inspection of the figure serves to demonstrate clearly that the frequency distribution of the metallic phase is entirely unlike the distribution of any major constituent thus far encountered in meteorites. Instead of one peak, as is usually the case, the metal-phase frequency curve is spread over a wide range of concentrations, within which three peaks can be rather clearly discerned. The first peak (0–1 per cent metal) and the second peak (7–10 per cent metal) appear definite. Statistical treatment indicates that the third peak (17–19 per cent metal) has a reasonably high probability of being definite. It is noteworthy that the valleys in between the maxima coincide with the discontinuities in the combined iron–metallic iron ratio depicted in figure 9.

If we assume that the meteorites studied are statistically representative of the silicate shell of the disrupted planet and if we assume, in addition, that the metal-phase content of the silicate shell increases with depth below the surface, then it becomes possible to obtain some idea of gross composition as a function of depth. Any precise statement of composition as a function of depth depends, of course, upon detailed knowledge of density as a function of pressure and of the weight of the core of the planet relative to the weight of the silicate mantle. As neither of the quantities can be expressed at present with any degree of exactness, we must confine our discussion for the time being to relative volumes, neglecting compressibility. In this basis, figure 11 has been transformed into a curve showing metal-phase content as a function of "fractional volume increments of the silicate mantle at zero pressure." Inspection of figure 12 serves to demonstrate that the increase in metal content with increasing depth is variable, sometimes being small and at other times being large, and that definite discontinuities exist. It is to be stressed that these discontinuities lie quite close to the discontinuities observed in the combined iron-metallic iron ratios, a curve compiled upon an entirely different basis.

It is, of course, most tempting to connect the metal-phase variations with the well-known seismic discontinuities of second order existing within the earth. If, as the meteorite data indicates, a planet can

[7] Fifty-seven analyses taken from O. C. Farrington's compilation (1911), in which the amount of metal phase had been determined with sufficient accuracy, were added to the 127 analyses comprising table 8.

possess such large metal concentrations and marked gradients of metal content, then one might expect the metal content to play a major role in determining seismic wave velocities. Any attempt, however, to correlate seismic discontinuities and metal-phase contents accurately must await further theoretical treatment of the density of matter under the pressures encountered in bodies of planetary dimensions.

If one assumes that the planet from which meteorites came possessed a core-to-mantle weight ratio comparable to that existing within the earth and if one further assumes that the differentiations within the planet were similar to those now existing within the earth, then certain comparisons can be made: (1) A metal-phase concentration of approximately 0.2 per cent by weight is attained at a depth of 0.02 radial units (130 km.). It is to be noted in this connection that free iron frequently has been observed to exist in certain basalts. (2) The maximum combined iron concentration is reached at a depth of about 0.07 radial units (450 km.). (3) The marked changes in be-

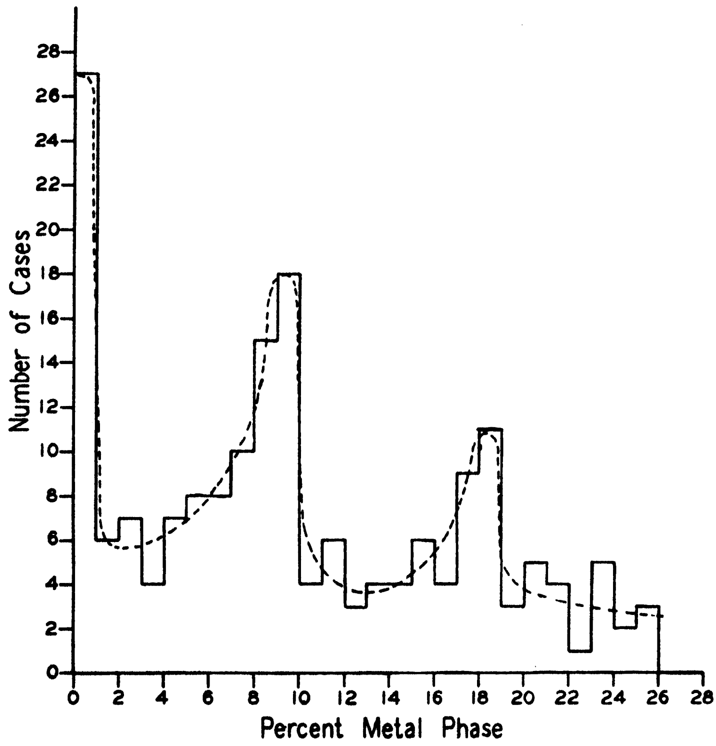



FIG. 11.—Frequency distribution of the metal phase of stony meteorites

havior of magnesium, silicon, calcium, and aluminum occur at a depth of about 0.07 radial units (450 km.). (4) The metal-phase curve discontinuities and the combined iron–metallic iron ratio discontinuities occur at depths of 0.08 radial units and 0.2 radial units (500 km. and 1,300 km.).

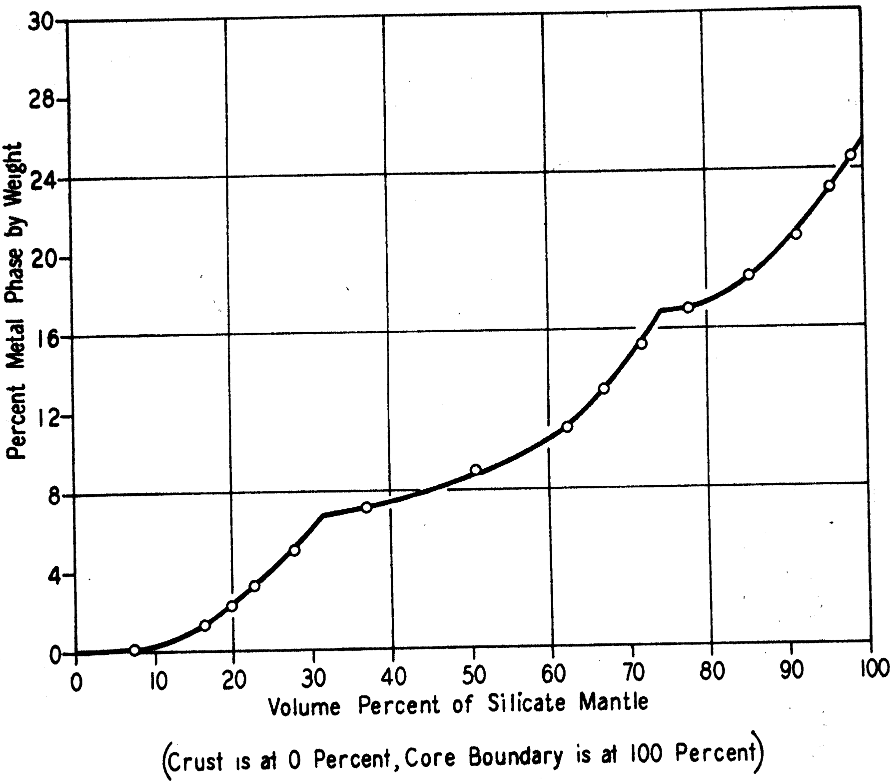



FIG. 12.—Metal-phase content of meteoritic planet as a function of volume percentage increments of the silicate mantle.

Before one takes such figures too seriously, however, it must be emphasized that, if pressure effects are as important as our results lead us to believe, then the chemical differentiations due to pressure should vary from planet to planet (considering only the minor planets) in such a way that the larger the planet, the greater the change in oxidation state with increasing depth. This naturally would result in differences in the relative locations of any discontinuities that might exist.

Summarizing the data concerning planet structure that have arisen from the present study of meteorites, the following can be said:

The meteorite parent-planet (and similarly the earth) was at one time quite hot—in all probability, actually molten. With its composition fixed by the fundamental abundances of the elements involved and the physicochemical processes involved in the initial formation of

the planet, pressure effects produced segregations of matter within the body. The ratio of combined iron to metallic iron for the planet as a whole was fixed by the reducing conditions in existence during the very beginning stages of the formation of the planet. However, owing to the large ionic radius of oxygen (small mass) and to the large density of metallic iron, shifts in the combined iron–metallic iron equilibrium were produced within the body in such a way that the greater the distance from the core, the greater the state of oxidation. As iron was the only major constituent of the magma that could readily give up oxygen, it played a major role in determining the states (combined or metallic) of the minor constituents of the planet. Thus a given metal probably distributed itself between the combined and free states according to the relationship

$$\frac{(M)_{si}}{(M)_{m}} = C\frac{(\mathrm{Fe})_{si}}{(\mathrm{Fe})_{m}},$$

where the ratio of combined iron to metallic iron was determined by the conditions outlined above. The constant $C$ depended upon the prevailing temperature, the thermal properties of the combined and uncombined forms of the minor constituent, the relative volumes of the combined and uncombined constituent, and the pressure. These considerations governed the structure of the planet from the core outward to the point where fractional crystallization effects near the surface produced further changes of substantial magnitude, detailed discussion of which would be beyond the scope of this paper.

## REFERENCES CITED


Barth, T. F. W. (1948) The distribution of oxygen in the lithosphere: Jour. Geology, vol. 56, pp. 41–49.

Brown, Harrison, and Patterson, Claire (1947*a*) The composition of meteoritic matter. I. The composition of the silicate phase of stony meteorites: Jour. Geology, vol. 55, pp. 405–411.

——— — ——— (1946*b*) The composition of meteoritic matter. II. The composition of iron meteorites and of the metal phase of stony meteorites: *Ibid.*, pp. 508–510.

Clarke, F. W., and Washington, H. S. (1922) The average chemical composition of igneous rocks: Nat. Acad. Sci. Proc., vol. 8.

Daly, R. A. (1933) Igneous rocks and the depths of the earth, New York: McGraw-Hill Book Co.

Farrington, O. C. (1911) Analyses of stone meteorites: Field Mus. of Nat. History Pub. 151, Geol. Ser., vol. 3, no. 3.

Goldschmidt, V. M. (1927) Krystallbau und chemische Zusammensetzung: Ber. vol. 60.

——— (1935) Grundlagen der Quantitaven Geochemie. II. Seltene Elemente in Meteoriten: Fortschr. Mineralogie, Kristallographie, Petrographie, vol. 19.

——— (1938) Geochemische Verteilungsgesetze der Elemente, Oslo.

Ishibashi, M. (1931) Über den bei der röntgenspektroskopischen Analyse auftretenden Kathodenstrahleneffekt: Zeitschr. anorg. Chemie, vol. 202.

Lewis, G. N., and Randall, Merle (1923) Thermodynamics and the free energy of chemical substances, New York: McGraw-Hill Book Co.

Neuburger, M. C. (1936) Gitterkonstanten für das Jahr 1936: Zeitschr. Kristallographie, vol. 93.

Noddack, I. and W. (1930) Die Häufigkeit der chemischen Elemente: Naturwiss., vol. 35.

——— (1934) Die geochemischen Verteilungs-Koeffizienten der Elemente: Svensk kemisk Tidsskr., vol. 46.

Pauling, L. (1940) Nature of chemical bond, 2d ed., Cornell University Press.

Prior, G. T. (1916) On the genetic relationships and classification of meteorites: Mineralog. Mag., vol. 18.

——— (1920) The classification of meteorites: *Ibid.*, vol. 19.

Richardson, W. A., and Sneesby, G. (1922) Frequency-distribution of the major oxides in the analysis of igneous rocks: Mineralog. Mag., vol. 19.

Roth, W. A., and Troitzsch, H. (1932–33) Arch. Eisenhüttenw., vol. 6.

Taylor, N. W. (1930) Die Kristallstrukturen der Verbindungen $Ni_2SiO_4$: Zeitschr. physikal. Chemie, vol. 9.

Wahl, W. A. (1910) Beiträge zur Chemie der Meteoriten: Zeitschr. anorg. Chemie, vol. 69.

Watson, F. G. (1939) The mean chemical composition of meteoritic accretion: Jour. Geology, vol. 47.

zur Strassen, H. (1930) Das Gleichgewicht zwischen Eisen, Nickel und ihren Silikaten im Schmelzfluss: Zeitschr. anorg. Chemie: vol. 191.

——— (1931) Die Wirkung von Schwefel aus das schmelzflüssige Gleichgewicht: *Ibid.*, vol. 200.

# STUDIES FOR STUDENTS

## A PREFACE TO THE CLASSIFICATION OF THE SEDIMENTARY ROCKS

F. J. PETTIJOHN
University of Chicago

"Perhaps if we stratigraphers insisted on a more refined classification of our sediments, instead of being satisfied with conglomerates, sandstones, shales, limestones, and some minor types, we would make more rapid progress; for it is my belief that precision in classification leads to precision in thought, and so is of vast value as a mental discipline. It might not be amiss to insist that a firm foundation in the classification of our rocks is a needful preliminary to the building of a permanent superstructure, and to urge that we get together and follow the lead of the pyro-petrographers."—A. W. GRABAU (1917).

### INTRODUCTION

A great deal of the confusion which pervades present-day thought and usage respecting the textures and structures, the nomenclature, and the classification of sedimentary rocks springs from failure to understand the fundamental character of these deposits and from an inadequate concept of the aim and objectives of rock nomenclature and classification. It seems advisable, therefore, to attempt to analyze and state the problems involved, in the hope that the paths which lead to a theoretically sound and practically useful conclusion will become evident.

This essay, as well as the papers by Shrock and Krynine which follow it, is an attempt to reopen the problem of classification of the sedimentary rocks, in the hope that a sound and workable system may be devised.

### BASIS OF ROCK CLASSIFICATION

The first strait jacket in which our thinking has been confined is the time-honored division of all rocks into the three categories: igneous, sedimentary, and metamorphic. This classification has, more than any other one thing, blocked constructive thinking on the classification problem. Grabau (1904) has been almost the only worker to try to break this mental log jam. His effort, though laudable, was bogged down by his attempt to introduce simultaneously an extended and unfamiliar set of rock names. Nonetheless, his recognition of two fundamentally different rock classes, named by him "exogenetic" and "endogenetic," is a sound contribution to the problem. Loosely stated, the exogene rocks are the clastic or detrital rocks; and into this category fall the familiar clastic deposits, such as tuff and sandstone. The endogenetic rocks are, in the vernacular, the "chemical" rocks—precipitates from solution (in the main)—such as rock salt, granite, and the like.

Each of the two main rock groups has its own characteristic textures and structures. Those of a tuff and a sandstone are much more alike than are those of sandstone and rock salt. On the other hand, the textures and structures of granite and rock salt are more akin than are those of tuff and granite. The reasons are self-evident. The clastic rocks have a similar

origin. The principles of fluid mechanics governing the transport and deposition are the same, irrespective of the mineral character of the material deposited. Therefore, the textures and structures of all exogene rocks are much alike. Similarly, the precipitation of materials from solution, governed by phase-rule chemistry, leads to similarity of textures and structures. Differences of temperature and composition of solutions may lead to different mineral composition, but they do not lead to significant differences in texture or structure. The similarity in character of rock salt and granite extends even beyond texture and structure. Invasive or piercement relations are shown by each. This may be more than a coincidence.

Either group, "clastic" or "chemical," may undergo reorganization after deposition. Such reorganization is metamorphism in the broadest sense. Metamorphic recrystallization gives rise to new attributes—the crystalloblastic textures and structures or those induced by recrystallization in the *solid state*. If these changes take place at relatively low temperatures and pressures, they are usually called "diagenetic." Nonetheless, the textures and structures do not differ materially from those induced at higher temperatures or pressures (unless the latter are differential). The differences observed are only those of mineral composition.

### GENETIC VERSUS DESCRIPTIVE CLASSIFICATIONS

Much that is wrong has been said about the need for a descriptive classification, independent of a classification based on genesis. There can be no such classification worthy of consideration. Genesis is the ultimate aim of any study of rocks, and no descriptive classification—so called—can be worth much unless the characters used for the classification are meaningful or significant. And the only test for significance is whether they are or are not basic to understanding of origin. Where would the "hard-rock" petrologist be if he were to classify all the coarse-grained igneous rocks on the basis of color? Such terms as "red phanerite," "pink phanerite," "white phanerite," etc., are more or less meaningless. Yet the present-day student of the sedimentary rocks is content to divide the sandstones upon such a chromatic basis.

Genesis must and does permeate our classification. Let one try to describe or define an arkose and distinguish it from a granite without reference to origin. Even the major categories with which most classifications begin—namely, igneous, sedimentary, and metamorphic—are genetic. One must decide origin before one can apply the term "granite" or the term "arkose."

In biology the basis of all sound taxonomic work must be selection of significant characters for classificatory purposes and avoidance of irrelevant peculiarities. Not all organisms with wings, for example, should be grouped together. So, too, any classification of sedimentary deposits, to be usable, must be based on significant properties. It cannot be argued that we do not know what the significant properties are. These are well known. The prevalent confusion, however, stems from failure to recognize the most basic division of rocks into the exogenetic and endogenetic groups, as pointed out above. The significant properties of one group are *not* the significant properties of the other. Hence an attempt to treat all sedimentary rocks alike will and must fail. Thus to apply one set of textural terms to all carbonate rocks obscures, rather than elucidates, their natural history.

### THE "HYBRID" CHARACTER OF SEDIMENTS

From what has been said it might be assumed that a rock was either exogenetic or endogenetic. Many are both and therefore exhibit the textures of each. This arises from the fact that such rocks, notably the sandstones (and also certain limestones), contain both clastic elements and a chemically precipitated cement. The textural characteristics are, therefore, hybrid. This hybrid character, however, does not vitiate what has been said above. Instead, it emphasizes the need for recognizing the dual character of many sedimentary rocks. Superimposed on the textures due to mechanical and chemical sedimentation and commonly obscuring them may be those textures induced by recrystallization and replacement (diagenesis). The proper description and interpretation of a sedimentary rock is, therefore, a job of no mean proportions. Sediments may exhibit a textural pattern more complex than that of an "igneous" rock.

Peculiar to the sedimentary rocks and absent from the igneous rocks are the organoform textures and structures. In most cases these are subordinate to the clastic or chemical textures, but in a few rocks they are dominant.

### ROCK NAMES

The problem of a proper name for a rock is indeed a difficult one. The students of the sedimentary rocks have attempted to frame a system of nomenclature such that each rock name would be an abbreviated description of the rock (e.g., Grabau). They have also attempted to give a more precise definition of rock terms acquired from the prescientific era. To date, neither effort has been very successful.

One direction in which progress might be made is to follow the example of the igneous petrologist and use a single name, commonly of geographic (type-locality) derivation, for common, recurring rock types. Rather than try to devise names with various prefixes and suffixes, why not use a single term? To the student of igneous rocks each of the terms "rhyolite," "gabbro," or "shonkinite," for example, denotes a rock with a special combination of textures and minerals. Such terms as "arkose" and "graywacke" are comparable terms that are familiar to the student of sedimentary rocks. These can be extended, however, and a term such as "bradfordite" might become a synonym for that particular type of subgraywacke described by Krynine from the Bradford district in Pennsylvania. One commonly refers to "the 'Bedford' (Spergen) type of limestone." Why not "spergenite"?

### A QUANTITATIVE CLASSIFICATION

The trend in science is always toward greater precision. So also in petrology. Commendable efforts have been made in recent years to quantify the terms "gravel," "sand," "silt," etc. These have resulted in general acceptance of a few standard size-grades adaptable to clastic sediments. Less successful or complete have been the efforts to extend this quantification to the naming of the aggregate. Though the size-grade "sand" is now established,[1] there is no agreement about the term "sandstone." Must all the material be between $\frac{1}{16}$ and 2 mm.? Must the average fall in this range? What average: median, mode, or mean? And what kind of mean? Or must 50 per cent fall in this range? Or will some

[1] For North American geologists the Udden grades from $\frac{1}{16}$ to 2 mm. are considered sand. European usage is different. So also is the usage of engineers and pedologists.

other proportion be acceptable? No standard usage prevails. But such standards can be established and, no doubt, will be in the future.

### TEXTURE AND CLASSIFICATION

The importance of the concepts here outlined can best be made clear by reference to a specific example, namely, a consideration of the textures of limestones. Confusion arises from the fact that limestones do not form a petrographically homogeneous group.[2] They are in part exogenetic or clastic, in part endogenetic ("chemical" and "organic"), and in the main "metamorphic" derivatives of these basic types. Unless this polygenetic character is recognized, no progress can be made toward a description of their textures and structures, or toward a rational nomenclature and classification. To attempt, for example, to impose a single scale of size-grades on such a heterogeneous group of rocks leads only to further confusion. The limestones of clastic origin—which are very common, notwithstanding statements to the contrary—can best be treated as clastics, and the textural terms applied to clastics can readily be applied to these rocks. The Udden geometric grade-scale is appropriate for such deposits. The appropriateness of this (or any other geometric grade-scale) arises from the fact that it tends to symmetrize the size-distribution curves. The tendency of such distributions to be log normal, as shown by this symmetrization, suggests that some underlying law of sorting or grading is operative and that such a scale, therefore, is significant and appropriate. To apply such a scale of size-grades to limestone with chemical textures or to those with "metamorphic" (crystalloblastic textures) is an error. We have no a priori reasons to believe that a geometric grade-scale has any meaning in such cases.

The textures of chemical precipitates ("igneous" textures) are well known, and established practices of description and nomenclature of these textures should be followed. Likewise, the textures of the crystalloblastic rocks, so commonly seen in limestones, should be recognized and described as such.

[2] The term "limestone," therefore, is, at best, a field term or a term useful in trade or commercial circles.

### MINERAL COMPOSITION AND CLASSIFICATION

The mineral composition, like texture, is an important element in any classification scheme. But mineral composition is not wholly independent of texture, and an analysis of the factors affecting the composition must be understood if it is to be rationally used for classificatory purposes.

Minerals are both exogenetic and endogenetic in origin. The chief exogenetic or detrital constituents of a clastic sediment are rock fragments, stable and labile mineral grains from the source rock, and the secondary products of weathering.

The rock fragments of phaneritic or coarse-grained rocks cannot, by the nature of things, occur in fine-grained sediments. Therefore, they characterize only the conglomerates. The fragments of the aphanitic or fine-grained rocks, on the other hand, can occur in the rocks of medium grain size (sands) and do occur in large volume in certain types, notably the graywackes. By and large, however, the rock fragments increase in importance with increase in grain size.

Sand grains are derived principally from the phaneritic igneous rocks. Both stable and labile minerals are present in the sands, but they vary widely in the

ratio of one to the other. The monomineralic particles of phaneritic origin dominate in the sands but are present in subordinate amounts in both coarser and finer sediments.

The products of chemical decay are of ultra-fine grain. Hence they characterize the mudstones. Under certain conditions, however, they may be coagulated and be deposited with the sands (as in the graywackes).

In addition to the differences in mineral composition related to the texture of the sediment, there are differences in the mineral composition *in any given size-grade* related to the maturity of the sediment. Maturity may be defined as the measure of the approach of the sediment to the stable end-state. The latter is the ultimate state to which a sediment tends to evolve. A stable sand, for example, would consist almost exclusively of quartz, be exceedingly well sorted, and be highly rounded.

There are various indices of maturity, of which mineral composition is, perhaps, the most important. In the sands, for example, the quartz-feldspar ratio is a very appropriate index. Although this ratio is governed by a number of factors, the prime one is the rate of erosion and concomitant rate of deposition. Inasmuch as the respective rates are controlled by rate of elevation and subsidence, the feldspar content is an index of crustal instability or *tectonism*. Correlative compositional differences due to tectonism may be observed in both the gravel and the clays. The influence of tectonics on the sedimentary regimen has been emphasized in the United States chiefly by Krynine, whose three major families of sediments, designated by their principal arenite member—the orthoquartzite, graywacke, and arkose clans—are indicative of three principal tectonic and geomorphic stages of the geosynclinal cycle.

Superimposed on the compositional differences related to grain size and to tectonism are those due to postdepositional or epigenetic changes—both solution and precipitation. The roles of intrastratal solution and precipitation are just beginning to be understood. The former is responsible for selective loss of the original detrital constituents. The latter is the process to which the mineral cement is due. Least understood are the extensive changes in composition due to simultaneous solution and deposition, that is, the metasomatic replacements.

These epigenetic changes are not, however, wholly independent of the tectonic history of the deposit, as Smithson has pointed out. The sediments of the geosyncline undergo a higher grade of epigenesis than do those of other places. The metasomatic replacement of the limestones and other rocks and the formation of chert seem to be promoted by the rise of the geotherms accompanying geosynclinal downwarp.

Epigenesis is also a function of time. The probability of change is improved with increase in age of the deposit. Therefore, there is a changing lime-magnesia ratio in the limestones with decreasing age, as was early noted by Daly. Similarly, there is an increase in the number and kind of heavy minerals with decrease in age, as noted by several investigators. The ratio of carbonate to silica cement in sandstones of various ages seems also to be a function of time as a consequence of the replacement of the former by the latter.

## CONCLUSIONS

Sediments are the product of clastic sedimentation (allogenic), precipitation (biochemical and chemical—authigenic

restricted), and postdepositional solution and precipitation (epigenesis). Each factor modifies the textures, structures, and composition of the sediment in a profound manner and gives rise to distinctive rock types.

A classification scheme, therefore, must be based on parameters which are significant in terms of the origin of the rock under consideration. Inasmuch as the sedimentary rocks are polygenetic, a simple biaxial tabulation of minerals and textures, such as serves for the orthomagmatic igneous rocks, will not do for the sedimentary rocks. Because the mineral composition of the clastic sediments is governed by three factors (grain size, tectonism, and mineral stability after burial), the variations and resultant rock types must be shown by three axes.

The hybrid nature of sediments, expressed by the textural and mineralogical complex which the sedimentary rocks exhibit, the need for greater precision and quantification, and the need for definitive rock names impose further restrictions on one who would propose a classification of the sedimentary rocks.

REFERENCE CITED

Grabau, A. W. (1904) On the classification of sedimentary rocks: Am. Geologist, vol. 33, pp. 228–247.

# A CLASSIFICATION OF SEDIMENTARY ROCKS

ROBERT R. SHROCK
Massachusetts Institute of Technology

ABSTRACT

A simple field and laboratory classification of sedimentary rocks is proposed. It is based fundamentally on composition and texture, with the primary divisions determined by the mode of origin of the constituents. It is tripartite in nature, consisting of (1) a dominantly fragmental division, including conglomerates, sandstones, and shales; (2) a division represented by rocks which are partly fragmental and partly precipitated, including ironstones, silicastones, coal, limestones, and dolostones; and (3) a division of salinastones, which are dominantly precipitated but may possibly be fragmental. The several new terms in the classification were first proposed in a paper read before the Chicago meeting of the Geological Society of America in December, 1946 (Shrock, p. 1231).

## INTRODUCTION

The present classification and taxonomic nomenclature of sedimentary rocks are not altogether satisfactory, for several reasons. Inexperienced students of geology need a simple classification consisting of a few general categories to which they can assign the rock specimens that they collect in the field or store in the laboratory pending detailed study. Certain of these categories have names which have long been used by geologists, e.g., conglomerate, sandstone, shale, ironstone, coal, and limestone. These names are quite satisfactory for our purpose and are used in the proposed classification. Certain other terms, although in use for a long time, are objectionable for one reason or another, e.g., "breccia," "dolomite," "anhydrite," and "gypsum." New terms are proposed to replace these. There is no satisfactory general term for sedimentary rocks consisting largely of silica or for those composed dominantly of the saline minerals. "Silicastone" and "salinastone" are proposed to fill this need. Finally, certain terms have come to have such widely differing or indefinite meanings that they need redefinition, e.g., "graywacke" and "shale." These are discussed and redefined.

The primary purpose of this discussion is to develop a classification of sedimentary rocks simple enough for effective use in the field, where ordinary laboratory and microscope facilities are not available, and in laboratories and museums, where general categories are needed for storage and display purposes. The proposed classification is based primarily on the mode of formation of the particles or crystals—*fragmental* or *precipitated;* secondarily on the two fundamental properties of a sedimentary rock—*composition* and *texture*.[1] These three aspects of sedimentary rocks ordinarily can be determined in the field by observation with the aid of a hand lens and a pocketknife. They can be verified readily in a modern petrographic or sedimentological laboratory.

The complete proposed classification is tabulated on the next page (table 1) and should be consulted in reading the following discussion.

[1] In a prepared discussion to accompany a paper by C. A. Bays and S. H. Folk on "Developments in the application of geophysics to ground-water problems" (1944), P. D. Krynine states: "A rock has really only two basic, fundamental properties; composition and texture, meaning that a rock is made up of certain constituents (generally minerals) put together in a certain way."

TABLE 1

A CLASSIFICATION OF SEDIMENTARY ROCKS

| | NATURE OF SEDIMENTS | | SEDIMENTARY ROCKS | | |
|---|---|---|---|---|---|
| DOMINANTLY FRAGMENTAL | Angular particles more than 2 mm. in greatest dimension | Rubble composed of sharpstones | SHARPSTONE | CONGLOMERATE | |
| DOMINANTLY FRAGMENTAL | Rounded particles more than 2 mm. in greatest dimension | Gravel composed of roundstones | ROUNDSTONE | CONGLOMERATE | |
| DOMINANTLY FRAGMENTAL | Angular and rounded particles of rocks and minerals ranging in greatest dimension from 2 mm. to .06 mm. | Volcanic fragments = Tuff<br>Mixture of rock and mineral fragments<br>Quartz + Feldspar<br>Quartz + other minerals in large amount<br>Quartz + other minerals in small amount | TUFFSTONE<br>GRAYWACKE<br>ARKOSE<br>NORMAL<br>QUARTZOSE | SANDSTONE | |
| DOMINANTLY FRAGMENTAL | Rock and mineral particles ranging in greatest dimension from .06 mm. to .001 mm. and colloidal particles less than .001 mm. in greatest dimension | Volcanic ash<br>Silt particles — .06 to .001 mm.<br>Clay materials<br>Silt + Clay + Water = Mud | ASHSTONE<br>SILTSTONE<br>CLAYSTONE<br>MUDSTONE | SHALE | |
| PARTLY FRAGMENTAL | $Fe^{II}$ and $Fe^{III}$ compounds precipitated inorganically and organically as concretions, nodules and layers<br>Impurities commonly present in the layers | Iron concretions<br>Iron compounds + mud, silica, etc. | Concretionary<br>Precipitated | IRONSTONE | PARTLY PRECIPITATED |
| PARTLY FRAGMENTAL | Siliceous inorganic fragments less than .06 mm. in greatest dimension<br>Siliceous organic hard parts and their fragments<br>Silica precipitated as oölites, pisolites, etc.<br>Silica precipitated from suspensions and solutions | Inorganic fragments<br>Diatom frustules, radiolarian skeletons and sponge spicules<br>Siliceous concretions<br>Chert, flint, sinter, etc. | Fragmental<br>Concretionary<br>Precipitated | SILICASTONE | PARTLY PRECIPITATED |
| PARTLY FRAGMENTAL | Plant structures — spores, fronds, leaves, wood, etc.<br>Inorganic sediment<br>Waxes, resins, etc. from decomposition of plants | Plant debris; inorganic impurities<br>Plant fluids | COAL | COAL | PARTLY PRECIPITATED |
| PARTLY FRAGMENTAL | Calcite and Aragonite fragments<br>Calcareous organic hard parts — shells, exoskeletons, plates, spines, and fragments<br>Organically and inorganically precipitated concretions<br>Inorganically precipitated $CaCO_3$ — Evaporation, etc.<br>Organically precipitated $CaCO_3$ — (1) by $NH_3$ from decomposition; (2) loss of $CO_2$ to plants; etc. | | Fragmental<br>Concretionary<br>Precipitated | LIMESTONE | PARTLY PRECIPITATED |
| PARTLY FRAGMENTAL | Dolomite fragments<br>Dolomitized organic hard parts<br>Dolomitic concretions<br>Inorganically precipitated dolomite<br>Organically precipitated dolomite | | Fragmental<br>Concretionary<br>Precipitated | DOLOSTONE | PARTLY PRECIPITATED |
| POSSIBLY FRAGMENTAL | Fragments of anhydrite, gypsum, halite, alkali, nitrate caliche, etc. | | Fragmental | SALINASTONE | DOMINANTLY PRECIPITATED |
| POSSIBLY FRAGMENTAL | Evaporites — minerals precipitated during evaporation of saline waters | Anhydrite<br>Gypsum<br>Chlorides<br>Nitrates<br>Other rare salts — Precipitated | ANHYDROCK<br>GYPROCK | SALINASTONE | DOMINANTLY PRECIPITATED |

## DOMINANTLY FRAGMENTAL SEDIMENTARY ROCKS

### INTRODUCTION

Fragmental sedimentary rocks are composed of discrete sedimentary particles, packed together in such manner as to give the mass some coherence. They usually contain more or less interstitial material, which acts as a binder.

The fragments range in size from large blocks many meters in greatest dimension to tiny particles of colloidal size. They vary in shape from irregular sharp-edged fragments, at the one extreme, to spheroidal and ellipsoidal grains, at the other. They may be fragments blown out of volcanoes and related vents[2] or mineral and rock fragments resulting from the disintegration of pre-existing rocks.

In some rocks the interstitial material merely coats the grains around their contacts; in others it partly or completely coats the grains but does not fill all the interstitial voids; in still others it fills the interstices to such an extent that porosity is reduced almost to zero (Krynine, 1941*e*, pp. 108–116). Ordinarily, interstitial material does not enter importantly into the problem of classification. However, if it constitutes greater volume than is necessary for cementation, as in some sandstones, or if it is a major rock constituent, as in coals and shales, it must be given appropriate recognition in defining or describing the rock.

Since, by definition, the rocks of this category are composed of particles or fragments, it follows that the original sediments were granular and could, on deposition, acquire certain features unique to granular materials. The possession of such features (e.g., cross-lamination and ripple-mark) precludes the possibility of the sediments' having been precipitated from solution. This same statement holds for cross-laminated or ripple-marked limestones and silicastones.

### CONGLOMERATES

The term "conglomerate" is here used for the rock resulting from a natural binding-together of rubble or gravel. Rubble is any natural accumulation of *angular* mineral, rock, or fossil fragments; gravel, any similar deposit of *rounded* fragments of the same nature (Wentworth, 1935, pp. 225–246). The minimum size of a fragment in these categories is arbitrarily taken as 2 mm. (Wentworth, 1922, pp. 377–392).

*Sharpstone conglomerate.*—Rubble particles may appropriately be designated "sharpstones." A sedimentary accumulation of such fragments, if bound together strongly enough to form a coherent mass, constitutes a "sharpstone conglomerate." If the interstitial material is as strong as the constituent particles, so that the rock breaks across it and the particles alike, the rock may be considered metamorphosed and designated "sharpstone conglomerite."[3] Acceptance of this nomenclature for sharpstone deposits will obviate further use of

[2] All volcanically ejected materials cooled sufficiently so as not to adhere to adjacent particles on reaching the earth's surface are considered sedimentary. Deposits composed of volcanic particles that are welded together are not considered sedimentary. The reader will find excellent discussions of the several kinds of pyroclastic deposits in two recent papers: (1) C. K. Wentworth and H. Williams (1932, pp. 19–53) and (2) F. G. H. Blyth (1940, pp. 145–156).

[3] B. Willard (1930, p. 438) proposed the term "conglomerite," defining it as follows: "It is suggested, therefore, that the term conglomerate be restricted to those pebbly rocks which break through the matrix and around the pebbles after the manner of sandstones. For the type in which fracture is through the pebbles and matrix, analogous to the conditions observed in quartzite, the term *conglomerite* is proposed."

"breccia," a term which now has several different meanings (Bonney, 1902, pp. 185–206; Norton, 1917, pp. 160–194; Reynolds, 1928, pp. 97–107).

*Roundstone conglomerate.*—The term "roundstone" was proposed by Fernald (1929, p. 240)

> . . . as a generic term to include the four largest sizes in Wentworth's schedule, boulder, cobble, pebble, and granule [+256 mm., 64–256 mm., 4–64 mm., and 2–4 mm., respectively] . . . [and] . . . to designate the unassorted accumulations composed of two or more sizes of rounded stones that occur in many situations.

Roundstones, therefore, can be thought of as the components of gravel, in the same way that sharpstones are the components of rubble. It is suggested that the rock composed of naturally cemented roundstone gravel be designated "roundstone conglomerate," and its metamorphosed equivalent, "roundstone conglomerite."

*Types of conglomerates.*—The student has only to observe the shape of the particles of a conglomerate to determine which of the two types it is. He can then use noun and adjectival modifiers to define the rock more exactly. Qualifying terms are available for expressing almost any aspect of the rock. Thus there are chert sharpstone conglomerates, quartzite roundstone conglomerates, and conglomerates described as basal, edgewise, desiccation, intraformational, coralline, shell, etc. Many other types have been described (Grabau, 1904, pp. 228–247; Mansfield, 1907, pp. 550–555; Field, 1916, pp. 29–66; Barrell, 1925, pp. 279–342; Twenhofel, 1936*b*, pp. 677–703; Allen, 1936, pp. 18–47; Pettijohn, 1943*b*, pp. 387–397), and the student will usually find, upon consulting the literature, that some geologist has previously described a conglomerate similar to the one he is studying.

### SANDSTONES

The term "sand" has long been used for any sedimentary accumulation of discrete mineral, rock, and fossil particles of a certain size—2–0.06 mm. in greatest dimension—regardless of their composition or mode of origin. Thus there are olivine, magnetite, glauconite, and quartz sands, for example; there are also volcanic sands, oölitic sands, coral sands, and foraminiferal sands. Many others have been enumerated (Allen, 1936, pp. 18–47; and Smith, 1946, pp. 121–143). Suffice it to emphasize that the term "sand" has a broad connotation rather than being limited to specific kinds of particles.

It would be logical to define a sandstone as a rock composed of sand, but such a broad definition would not be acceptable to most geologists (e.g., a rock composed of foraminiferal shells or calcareous hard parts of other animals would be called a "limestone," even though the particles of which it is composed would have been designated "sand" when they lay on the beach or on a shallow bottom). Therefore, "sandstone" is here used in a somewhat restricted sense for a sedimentary rock composed dominantly of quartz grains, feldspar grains, or bits of rock or mixtures of these, with minor amounts of other minerals.

For the present purpose five types of sandstone are included in the proposed classification. These are designated by second-order terms: "tuffstone," "graywacke," "arkose," "normal sandstone," and "quartzose sandstone." As with other sedimentary rocks, noun and adjectival modifiers can be used when necessary to define the nature of a sandstone more fully (e.g., glauconitic sandstone and micaceous sandstone). The reader will find an excellent discussion of sand-

stones in the previously cited article by Allen (1936, pp. 18–47).

*Tuffstone.*—Sedimentary accumulations of volcanic fragments of sand size might well be designated "tuffstone" and considered a special type of sandstone. There are many tuffstones of this nature in the geologic section, especially in the Pre-Cambrian, and they deserve a specific rock name (Bailey, 1926; Wentworth and Williams, 1932, pp. 19–53, Blyth, 1938, pp. 392–404; 1940, pp. 145–156).

*Graywacke.*—The term "graywacke," which has been much misused (Krynine, 1941*a*, pp. 2071–2074), may be simply defined as *a sandstone composed of a mixture of rock and mineral fragments ranging in greatest dimension from 2 to 0.06 mm.* This definition conforms with the classical meaning of the term, as well as with preferred modern usage. Knopf (in Pirsson and Knopf, 1926, p. 340), for example, defines the term as follows:

> *Graywackes.*—These are sandstone-like rocks of a prevailing gray color, sometimes brown to blackish, which, in addition to quartz and feldspar of an arkose, contain rounded or angular bits of other rocks, such as fragments of shale, slate, quartzite, granite, felsite, basalt, etc., or of varied minerals, hornblende, garnet, tourmaline, etc.

This definition, as well as the simpler one stated earlier, satisfactorily describes the Third Bradford sand of Pennsylvania, which Krynine (1940) considers a typical graywacke.

Graywackes are of great importance the world over in the earlier pre-Cambrian sedimentary record, sharing with arkoses a dominance over normal sandstones (Pettijohn, 1943*a*, pp. 925–972). If the term "graywacke" is to have significance with reference to geosynclinal deposition—and this is highly desirable (Jones, 1938; Krynine, 1941*c*, p. 1916)—then it should be used in its classical sense.

*Arkose.*—An arkose is a sandstone containing a relatively high percentage of feldspar grains (i.e., 25–50 per cent by weight). Many sandstones contain a few scattered particles of feldspar; arkoses, however, are composed dominantly of the two minerals quartz and feldspar. The term "arkose," as used by many geologists, also has genetic significance (Krynine, 1941*b*, pp. 1918–1919).

*Normal sandstone.*—The term "normal" is applied to those sandstones composed dominantly of quartz but with large amounts of other minerals. Normal sandstones are widely represented in the geologic section from the oldest to the youngest rocks. They exhibit considerable variation in the nonquartz components and in some cases contain small amounts of rock particles in addition to the mineral grains. They lie between graywackes, on the one hand, and quartzose sandstones, on the other. The typical normal sandstone does not have a conspicuous amount of any mineral other than quartz; some, however, are characterized by large amounts of nonquartz minerals (e.g., the glauconitic sandstones of the Wisconsin Upper Cambrian [Twenhofel, 1936*a*, pp. 472–487] and of the New Jersey Cretaceous). In the latter case the appropriate noun or adjectival modifier can be used to indicate the abundant nonquartz mineral (e.g., glauconite or glauconitic sandstone and micaceous sandstone).

*Quartzose sandstone.*—The term "quartzose" is applied to those sandstones composed almost entirely of quartz grains. The nonquartz minerals are typically small in amount (generally less than 1 per cent) and restricted to a few durable species (e.g., zircon, tourmaline, garnet, and ilmenite). Typical representatives of this type of sandstone are the St. Peter and Oriskany. Quartz-

ose sandstones are composed of multicycle sand grains representing sediments that have been eroded and deposited repeatedly. As a consequence, the grains are frosted in some rocks and are commonly well rounded.

*Metamorphosed sandstones.*—Silica-bound quartzose sandstones in which fracture crosses interstitial silica and quartz grains alike are designated "quartzites." Several kinds of quartzites are now generally recognized (Krynine, 1941*d*, pp. 1915–1916). "Arkosite" (Gruner, 1941, pp. 1577–1642; and Pettijohn, 1943*a*, pp. 925–972) has been applied to metamorphosed arkoses of the Lake Superior pre-Cambrian; and "graywackite," if it has not already been proposed, might well be used for metamorphosed graywackes.

SHALES

Shale is a laminated argillaceous rock, composed of a complex of silt fragments, colloidal micelles (which consist in a dispersed medium of a particle ranging in maximum dimension from 1 $\mu$ [0.001 mm.] to 1 m$\mu$ [0.000,001 mm.], adsorbed ions, and bonded water), and impurities such as organic materials. The rock is sufficiently consolidated and lithified that it maintains its essential character when weathered (especially when wetted after having been dried out).[4] Inasmuch as shales vary greatly in degree of consolidation and lithification, in the relative amounts of constituent materials, and in the development of lamination, to say nothing of their content of sand grains and organic matter, it seems advisable, and, in fact, it has now become rather common practice, to recognize by name three distinct types of subshale rocks—*siltstone*, *claystone*, and *mudstone* (Twenhofel, 1937, pp. 81–106). In addition, sedimentary deposits of volcanic ash deserve a special designation, and the term "ashstone" is suggested. Although "colloidstone" has been proposed (Alling, 1943, p. 266) for sedimentary rock composed of particles smaller than 1 $\mu$ (0.001 mm.), that term is not used in the proposed classification because the size range that it covers is included in the definition of shale.

*Ashstone.*—The geologic section contains many sedimentary deposits of volcanic ash. Such rocks are important and distinctive enough to be given a special designation. "Ashstone" is suggested for rock composed of particles of volcanic ash less than 60 $\mu$ (0.06 mm.) in greatest dimension. (Some writers would prefer a smaller figure.) Several terms have already been proposed for the finer-grained pyroclastics, but these apply more appropriately to recent accumulations (Wentworth and Williams, 1932, pp. 19–53; and Blyth, 1938, pp. 392–404; 1940, pp. 145–156).

*Siltstone.*—The term "siltstone" is in common use for indurated silt (Wentworth, 1922, p. 381). It is composed chiefly of mineral, rock, and fossil fragments with maximum dimensions from 0.06 ($\frac{1}{16}$ mm.) to 0.0039 mm. ($\frac{1}{256}$ mm.). The definition of siltstone given above follows general usage with respect to the limiting dimensions of the constituent particles (i.e., 0.06–0.0039 mm.). The author, however, prefers to extend the smaller dimension to 1 $\mu$ (0.001 mm.) in order to include particles now commonly referred to the so-called "clay size." Inasmuch as particles between 0.004 and 0.001 mm. in greatest dimension can be of some mineral other than a clay mineral (e.g., quartz), it would seem advisable to eliminate the term "clay" as a size

[4] It is pointed out later that mudstone, which may be considered a type of subshale, slakes to mud when dried out and wetted repeatedly.

designation. Furthermore, since 1 $\mu$ is the wave length for visible light and has been taken by several investigators as the upward limiting dimension for a colloidal particle, it is a logical dimension to use as a downward limit for silt particles. Therefore, accepting the suggested modification of the definition of silt not only would obviate further use of clay as a size term but would also fix the downward limit of a silt particle at the beginning of the colloidal range. This modification is incorporated in table 1. Siltstone feels harsh to the touch and does not lose its coherence when dried out and wetted repeatedly.

*Claystone.*—"Claystone" has had wide usage for very fine-grained, somewhat unctuous, conchoidally fracturing sedimentary rock composed largely of clay material (Grim, 1942, pp. 225–275). Claystone spalls when weathered, but it does not lose its solidity, i.e., it does not become mud.

*Mudstone.*—"Mudstone" is an appropriate term for those partly indurated argillaceous rocks which slake readily to mud when dried out and wetted repeatedly. The term has been in use for over a century, but not always with the connotation just given (Twenhofel, 1937, p. 90).

*Designation and description of shales.*—Ashstone, siltstone, claystone, and mudstone may all be regarded as partly lithified shales. Therefore, if they cannot be identified certainly in the field, they can be designated "shale" until their exact nature can be established in the laboratory.

All sorts of modifiers are available for indicating the many characteristics of argillaceous rocks. A few examples are ferruginous claystone, carbonaceous mudstone or shale, black shale, sandy (=quartz) siltstone, pencil and paper shales, oil shale, and calcareous shale.

*Metamorphosed shales.*—Argillite, slate, phyllite, and schist are members of a series of metamorphosed argillaceous materials. The sedimentary origin of argillites and slates is made evident by such features as graded bedding and coarse sandy layers. Phyllites and schists, on the other hand, rarely possess inherent features by which it can be certainly established that they were derived from sedimentary rocks.

### TILLSTONE

The few deposits of ancient glacial till that have been recognized as such are usually referred to as "tillite." It would make for more uniform classification if "tillstone" were adopted for consolidated and "lithified till" and "tillite" were reserved for metamorphosed tillstones. This modification of terminology should not cause too much disturbance, inasmuch as some of the better-known ancient till deposits (e.g., Gowganda, Dwyka, and Squantum) are more or less metamorphosed, hence are tillites (Coleman, 1926). "Tillstone" is not included in the proposed classification because it implies genesis. A tillstone would fall in one or the other of the two categories of conglomerate.

## PARTLY FRAGMENTAL AND PARTLY PRECIPITATED ROCKS

### INTRODUCTION

The rocks included in this indefinite subdivision are alike in being composed of mixtures of organic and inorganic fragments or concretions and precipitated compounds.

The particles may be fragments resulting from the breaking-up of organically secreted hard parts or structures (e.g., broken and comminuted corals);

they may be skeletal or shell fragments or the skeletons and shells themselves, released upon decomposition of enclosing organic tissue (e.g., sponge spicules, echinoid plates, radiolarian skeletons, foraminiferal tests, and diatom frustules); or they may be concretions (e.g., oölites and pisolites), built in several different ways. They may, on the other hand, come from an inorganic source (e.g., siliceous inorganic materials resulting from the weathering of chert and cherty limestones and included under the term "tripoli" [Metcalf, 1946, pp. 1–25]).

The precipitated parts of the rocks were produced by organically or inorganically induced flocculation of colloids and precipitation of soluble salts. These substances usually form the matrix of the rock, filling the voids between fragments and concretions (e.g., radiolarian silicastone and oölitic silicastone). They constitute the entire rock in some limestones and dolostones and in many silicastones.

Suffice it to point out that the five types of sedimentary rocks included in this general category show considerable range in composition and texture and in the proportions of fragmental and precipitated constituents. It will be necessary, therefore, as in the previous types, to employ descriptive modifiers to indicate the special characteristics of any general rock type.

### IRONSTONES

The term "ironstone" has commonly been applied to thin beds of hard, tough, iron-bearing argillaceous rock that is characteristic of coal-bearing sequences the world over. It might also be applied to other sedimentary rocks in which compounds of iron are a major constituent (e.g., a lateritic iron ore, such as that lying between Miocene lava flows in Oregon [Williams and Parks, 1923, pp. 1–44]). Oölitic iron ores of the Clinton type represent concretionary ironstones.

### SILICASTONES

"Chert," "flint," "hornstone," and "novaculite" are familiar terms for certain siliceous rocks. "Diatomite" and "radiolarite" have been applied to rocks composed of the siliceous frustules of diatoms and the siliceous skeletons of radiolaria, respectively. There are certain siliceous rocks composed largely of concretions (mainly oölites) and others, which are rare, that are made up of very fine silica fragments (tripoli) derived originally from the chemical weathering of chert, cherty limestones, and novaculite. There is no satisfactory general term for this group of widely different, typically fine-grained siliceous rocks. "Silicastone" is suggested as such a general term. If it is desirable, the rock can always be classified more precisely, after necessary laboratory work, by using appropriate modifiers (e.g., cherty silicastone, diatomaceous silicastone, and oölitic or concretionary silicastone). In the field the more general term "silicastone" can be used when there is doubt concerning the true nature of the siliceous rock.

Much has been written about certain silicastones, and some controversy still exists over the origin of a few of them (Tarr, 1926, pp. 1–46; 1938, pp. 8–27; Tarr and Twenhofel, 1932, pp. 519–546).

### COALS

Coal is a complex of plant debris, solidified organic compounds derived from plant tissue, and inorganic impurities washed or blown into the coal swamps from the surrounding land areas (White and Thiessen, 1913, pp. 1–390;

Thiessen, 1947, pp. 1–53; Stutzer and Noé, 1940, pp. 1–461; Moore, 1940). It is unnecessary for the present purpose to discuss the several kinds or grades of coal. The exact nature of a coal can be determined only in a well-equipped laboratory. The beginning field geologist needs only the general term for his notes.

The proposed classification does not include deposits of solid hydrocarbons, such as asphalt, gilsonite, etc. These represent special types of sedimentary deposits and are exceedingly rare as compared with those included in the classification.

### LIMESTONES

The term "limestone," as it is now used almost universally in North America, includes three distinct kinds of calcareous rocks:

1. *Fragmental*, consisting of the fragments of calcareous fossils, of calcite crystals, and of broken concretions
2. *Concretionary*, composed of oölites, pisolites, stromatolites, biostromes, bioherms, and other masses of calcareous material built by organisms or formed inorganically
3. *Precipitated*, consisting of fine-grained calcareous material precipitated from solution either organically or inorganically

Fragmental limestones, the most abundant of all calcareous rocks, commonly exhibit typical sedimentary features of granular deposits, such as ripple-mark and cross-lamination. Concretionary limestones are likely to be local in extent, variable in thickness, and uneven in texture. Precipitated limestones, which are rare as compared to the two preceding types, are likely to be very fine grained, uniformly textured, commonly somewhat argillaceous, and in some cases mud-cracked (Tarr, 1925, pp. 252–264; Gee, 1932, pp. 162–166; Wood, 1941, pp. 192–200). Precipitated limestones lack all sedimentary structural features which depend upon a megascopic granular condition of the original sediments.

It is considered more desirable to use the term "limestone" in its broad sense than to propose new names for the three types just described. As in previous types, modifiers can be used to indicate special characteristics of the rock. A few such words are "chalk" and "marl" ("chalkstone" and "marlstone" might be appropriate rock names in certain cases), "travertine," "tufa," "coquina," "shell," "coralline," "algal," "dolomitic," "hydraulic," "lithographic," and "biohermal" or "reef."

Marbles, the metamorphosed equivalents of limestones (and dolostones), are of great variety because the calcareous rocks from which they were formed were widely different. It is worth while to emphasize that fragmental marbles come only from fragmental limestones. As with other metamorphosed sedimentary rocks, some marbles have been so completely recrystallized that any original sedimentary textures and structural features have been totally destroyed, making uncertain the nature of the original calcareous rock.

Certain limestones are recrystallized early in their history and have some of the properties of metamorphic marbles. These have been designated "diagenetic marbles."

### DOLOSTONES

The term "dolostone" is proposed for those sedimentary rocks consisting largely of the mineral dolomite, such as the Niagaran dolomitic rocks of Illinois (Willman, 1943, pp. 1–89). Adoption of this term would avoid the confusion arising from the use of a mineral name for both a mineral species and a rock type.

The presence in dolostones of other minerals and substances and of certain textural and structural features can be indicated by appropriate modifiers (e.g., calcitic, oölitic, and saccharoidal dolostone). Many rocks which have been described as dolomites might better be called "dolomitic" or "magnesian limestone" if dolomite is not the dominant mineral.

Dolomite marbles resulting from metamorphism of dolostones seem to be rare. Recrystallized dolostones, on the contrary, are not uncommon.

### DOMINANTLY PRECIPITATED AND POSSIBLY FRAGMENTAL ROCKS

This small subdivision includes those saline rocks usually called "evaporites." "Salinastone" is suggested as a general term for this group of sedimentary rocks.

The commonest salinastones are those composed of gypsum and anhydrite. "Gyprock" has already found use for the former, and "anhydrock"[5] is suggested as a suitable term for the latter. The composition of other less common salinastones can be indicated by appropriate modifiers (e.g., halite salinastone and borate salinastone).

### SUMMARY

The classification of sedimentary rocks here suggested is based primarily on whether the constituents are fragmental or precipitated and secondarily on composition and texture. Both sets of properties can usually be determined readily in the field by direct observation and by simple tests. Texture and composition are the basis of the general group terms (sharpstone and roundstone conglomerates, sandstone, shale, ironstone, silicastone, coal, limestone, and dolostone), which are names of the first order. Distinct subtypes (e.g., tuffstone, graywacke, arkose, ashstone, siltstone, claystone, and mudstone) bear names of the second order. Noun and adjectival modifiers constitute terms of the third order. It should be possible to assign a first-order name to any common sedimentary rock encountered in the field. Second- and third-order names had best be reserved for use in the laboratory after the nature of the rock has been determined in greater detail.

Students will find the several categories of sedimentary rocks of the proposed classification fully described in such standard works as *Treatise on Sedimentation* (Twenhofel, *et al.*, 1932); *Sedimentary Petrography* (Milner, 1940); *A Handbook of Rocks* (Kemp, 1940); and *Rocks and Rock Minerals* (Pirsson and Knopf, 1947).

Recent studies of oil sands (Krynine, 1940), microlithologies (Alling, 1945, pp. 737–755), and textures of carbonate rocks (De Ford, 1946, pp. 1921–1928), to cite but three, are good indices of the rich field of research that lies ahead for the beginning student of sedimentary rocks. It is hoped that the classification here proposed will serve the student as a simple filing system, which he can elaborate and refine as the need arises.

ACKNOWLEDGMENTS.—The writer is indebted to many of his colleagues for valuable criticism and suggestions. Special acknowledgments are due Drs. W. L. Whitehead and H. W. Fairbairn, of Massachusetts Institute of Technology, and Dr. H. C. Stetson, of Harvard University.

[5] "Anhydrock" was suggested to the writer by his colleague, Professor W. L. Whitehead, of Massachusetts Institute of Technology. It seems to be the most euphonious word that can be formed from "anhydrite" and "rock."

## REFERENCES CITED


ALLEN, V. T. (1936) Terminology of medium-grained sediments (with notes by P. G. H. BOSWELL): Nat. Research Council Ann. Rept. 1935–1936, Appendix I, Rept. Comm. Sedimentation, pp. 18–47.

ALLING, H. L. (1943) A metric grade scale for sedimentary rocks: Jour. Geology, vol. 51, pp. 259–269.

———. (1945) Use of microlithologies as illustrated by some New York sedimentary rocks: Geol. Soc. America Bull. 56, pp. 737–755.

BAILEY, T. L. (1926) The Gueydan, a new Middle Tertiary formation from the southwestern coastal plain of Texas: Univ. of Texas Bull. 2645.

BARRELL, J. (1925) Marine and terrestrial conglomerates: Geol. Soc. America Bull. 36, pp. 279–342.

BAYS, C. A., and FOLK, S. H. (1944) Developments in the application of geophysics to ground water problems: Illinois Geol. Survey Circ. 108.

BLYTH, F.G.H. (1938) Pyroclastic rocks from the Stapley volcanic group at Knotmoor, near Minsterley, Shropshire: Geol. Assoc. Proc., vol. 49, pp. 392–404.

———. (1940) The nomenclature of pyroclastic deposits: Volcanologique Bull. 6, ser. 2, pp. 145–156.

BONNEY, T. G. (1902) On the relation of certain breccias to the physical geography of their age: Geol. Soc. London Quart. Jour., vol. 58, pp. 185–206.

COLEMAN, A. P. (1926) Ice ages, recent and ancient, New York, Macmillan Co.

DE FORD, R. K. (1946) Grain size in carbonate rock: Am. Assoc. Petroleum Geologists Bull. 30, pp. 1921–1928.

FERNALD, F. A. (1929) Roundstone, a new geologic term: Science, vol. 70, p. 240.

FIELD, R. M. (1916) A preliminary paper on the origin and classification of intraformational conglomerates and breccias: Ottawa Naturalist, vol. 30, pp. 29–66.

GEE, H. (1932) Inorganic marine limestone: Jour. Sedimentary Petrology, vol. 2, pp. 162–166.

GRABAU, A. W. (1904) Classification of sedimentary rocks: Am. Geologist, vol. 33, pp. 228–247.

GRIM, R. E. (1942) Modern concepts of clay minerals: Jour. Geology, vol. 50, pp. 225–275.

GRUNER, J. W. (1941) Structural geology of the Knife Lake area of northeastern Minnesota: Geol. Soc. America Bull. 52, pp. 1577–1642.

JONES, O. T. (1938) On the evolution of a geosyncline (presidential address): Geol. Soc. London Quart. Jour., vol. 94, pp. lx–cx.

KEMP, J. F. (1940) A handbook of rocks, revised and edited by F. M. GROUT, 6th ed., New York, D. Van Nostrand Co., Inc.

KRYNINE, P. D. (1940) Petrology and genesis of the Third Bradford sand: Pennsylvania State Coll. Min. Ind. Exper. Sta. Bull. 29.

———. (1941*a*) Graywackes and the petrology of Bradford Oil Field, Pennsylvania: Am. Assoc. Petroleum Geologists Bull. 25, pp. 2071–2074.

———. (1941*b*) Paleogeographic and tectonic significance of arkoses (abstr.): Geol. Soc. America Bull. 52, pp. 1918–1919.

———. (1941*c*) Paleogeographic and tectonic significance of graywackes (abstr.): *Ibid.*, p. 1916.

———. (1941*d*) Paleogeographic and tectonic significance of sedimentary quartzites (abstr.): *Ibid.*, pp. 1915–1916.

———. (1941*e*) Petrographic studies of variations in cementing material in the Oriskany sand: Pennsylvania State Coll. Min. Ind. Exper. Sta. Bull. 33, pp. 108–116.

MANSFIELD, G. R. (1907) The characteristics of various types of conglomerate: Jour. Geology, vol. 15, pp. 550–555.

METCALF, R. W. (1946) Tripoli: Dept. Interior, Bur. of Mines Inf. Circ. 7371, pp. 1–25.

MILNER, H. B. (1940) Sedimentary petrography, London, Thomas Murby & Co.

MOORE, E. S. (1940) Coal, its properties, analysis, classification, geology, extraction, uses, and distribution, 2d ed., New York, John Wiley & Sons, Inc.

NORTON, W. H. (1917) A classification of breccias: Jour. Geology, vol. 25, pp. 160–194.

PETTIJOHN, F. J. (1943*a*) Archean sedimentation: Geol. Soc. America Bull. 54, pp. 925–972.

———. (1943*b*) Basal Huronian conglomerates of Menominee and Calumet districts, Michigan: Jour. Geology, vol. 51, pp. 387–397.

PIRSSON, L. V., and KNOPF, A. (1926) Rocks and rock minerals, New York, John Wiley & Sons, Inc.

——— — ———. (1947) Rocks and rock minerals, 3d ed., New York, John Wiley & Sons, Inc.

REYNOLDS, S. H. (1928) Breccias: Geol. Mag., vol. 65, pp. 97–107.

SHROCK, R. R. (1946) Classification of sedimentary rocks (abstr.): Geol. Soc. America Bull. 57, p. 1231.

SMITH, E. R. (1946) Sand: Indiana Acad. Sci. Proc., vol. 55, pp. 121–143.

STUTZER, O., and NOÉ, A. C. (1940) Geology of coal, Chicago, University of Chicago Press.

TARR, W. A. (1925) Is the chalk a chemical deposit? Geol. Mag., vol. 62, pp. 252–264.

———. (1926) The origin of chert and flint. Univ. of Missouri Studies, vol. 1, no. 2, pp. 1–46.

———. (1938) Terminology of the chemical siliceous sediments: Nat. Research Council Ann. Rept. 1937–1938, Rept. Comm. Sedimentation, Exhibit A, pp. 8–27.

——— and TWENHOFEL, W. H. (1932) Chert and

flint, in Treatise on sedimentation, rev. ed., Baltimore, Williams & Wilkins Co., pp. 519–546.

THIESSEN, R. (1947) What is coal? Dept. Interior, Bur. of Mines Inf. Circ. 7397, pp. 1–53.

TWENHOFEL, W. H. (1936*a*) The greensands of Wisconsin: Econ. Geology, vol. 31, pp. 472–487.

———. (1936*b*) Marine unconformities, marine conglomerates, and thickness of strata: Am. Assoc. Petroleum Geologists Bull. 20, pp. 677–703.

———. (1937) Terminology of the fine-grained mechanical sediments: Nat. Research Council Ann. Rept. 1936–1937, Appendix I, Rept. Comm. Sedimentation, pp. 81–106.

——— *et al.* (1932) Treatise on sedimentation, Baltimore, Williams & Wilkins Co.

WENTWORTH, C. K. (1922) A scale of grade and class terms for clastic sediments: Jour. Geology, vol. 30, pp. 377–392.

———. (1935) The terminology of coarse sediments (with notes by P. G. H. BOSWELL): Nat. Research Council Bull. 89, pp. 225–246.

——— and WILLIAMS, H. (1932) The classification and terminology of the pyroclastic rocks: Nat. Research Council Bull. 89. pp. 19–53.

WHITE, D., and THIESSEN, R. (1913) The origin of coal, with a chapter on the formation of peat by CHARLES A. DAVIS: Dept. Interior, Bur. of Mines Bull. 38, pp. 1–390.

WILLARD, B. (1930) Conglomerite, a new rock term: Science, vol. 71, p. 438.

WILLIAMS, I. A., and PARKS, H. M. (1923) The limonite ores of Columbia County, Oregon: Oregon Bur. of Min. Geology, vol. 3, no. 3, pp. 1–44.

WILLMAN, H. B. (1943) High-purity dolomite in Illinois: Illinois Geol. Survey Rept. Inv. 90, pp. 1–89.

WOOD, A. (1941) "Algal dust" and the finer-grained varieties of Carboniferous limestone: Geol. Mag., vol. 78, pp. 192–200.

# THE MEGASCOPIC STUDY AND FIELD CLASSIFICATION OF SEDIMENTARY ROCKS

PAUL D. KRYNINE
Pennsylvania State College



## PART 1. DEFINITION OF BASIC TERMS AND FUNDAMENTALS OF COMPOSITION, TEXTURE, AND STRUCTURE

The necessary knowledge is that of what to observe.—EDGAR ALLAN POE.

### INTRODUCTION

The systems of classification of sedimentary rocks currently in vogue are either entirely too generalized or tend to combine several dissimilar features, generally texture (and, specifically, grade size), together with some other factor, such as bulk chemical composition or mode of genesis, which, regardless of their possible combinations, cannot result in an unequivocal classification. In-

deed, one of these factors may be either somewhat subjective (genesis) or at least not susceptible of objective and accurate description at the megascopic level (bulk composition). As a result, it is extremely difficult to translate some of the existing sedimentary names into quantitative, objective, and reproducible terms; and, conversely, many of the same names when employed by different workers may carry entirely different meanings.

At present the basis of sedimentary classification is somewhat arbitrary, with four main systems in general use. These classification systems, which may be combined in several ways, are as follows:

1. By mode of origin (very common)
   Clastic or fragmental rocks
   Chemical precipitates
   Biogenic products (plant and animal)
2. By medium in which rocks originate
   Aqueous (water-laid)
   Eolian (wind-laid)
   Glacial (ice-laid)
3. By bulk composition
   Arenaceous (sandy)
   Argillaceous (clayey)
   Calcareous (limy)—although many other chemical precipitates are frequently added to this scheme
4. By texture and grain size, with many variations, permutations, and the addition of much involved terminology
   Psephites (coarse clastic rocks)
   Psammites (medium clastic rocks)
   Pelites (fine clastic rocks)
   Crystalline (chemical rocks)
      Phanerocrystalline (coarse and medium)
      Aphanitic (fine and very fine)

None of these schemes can compare in completeness, simplicity, or objectivity with the mineral composition–texture scheme in use for the classification of igneous rocks.

As a result, considerable scientific confusion is found to prevail in published descriptions of sediments. Attempts by the writer to teach (mostly unsuccessfully) undergraduate students these different unco-ordinated schemes of sedimentary classification, after putting them through the simple and logical nomenclature of igneous petrography, showed only too well how such confusion is likely to develop.

This confusion is undesirable not only from the point of view of "pure science" but also from the applied, economic aspect. Indeed, it should be clear to everybody that, in the field of igneous, so-called "hard-rock" mining geology, an economic geologist who is not capable of differentiating a granite from a gabbro and who insists in lumping these two rocks together under the vague term of "phanerite" is not exactly an asset to any employer. In fact, such a man would not be able to hold his job for one single day.

However, in the current practice of petroleum geology all the medium-grained clastic rocks are usually lumped together under the term of "sandstone." It happens, however, that a successful search for stratigraphic traps requires, as a necessary working tool, a much more precise and refined terminology than such a vague and ill-defined catch-all term.

As a result of work on the geosynclinal sediments of the Appalachian region from 1937 on, the writer finds that there appears to be a series of relationships between the genesis, composition, structure, and texture of sediments. Some of these concepts were presented in preliminary form elsewhere (1942, pp. 537–561; 1943; 1945, pp. 12–22) and others will be published in a complete form in the near future.

In the meantime this work has led to the formulation of a simplified form of sedimentary rock classification based on mineral composition and texture which forms the subject of the present discussion. This classification has been used

with success at the Pennsylvania State College since 1942 for teaching advanced sedimentology to Seniors and graduate students and since 1944 for teaching elementary hand-specimen petrology to Sophomores.

The research work that led to the formulation of these concepts was financed under Research Projects E-6 and E-14 by the Mineral Industries Experiment Station of the Pennsylvania State College, Dr. A. W. Gauger, director. To his colleague, Dr. J. C. Griffiths, the author is indebted for several excellent suggestions that have been incorporated into this paper.

## THE MAKEUP OF SEDIMENTS

A rock, or any other solid for that matter, has only two basic, fundamental properties: composition and texture, meaning that a rock is made up of certain constituents (generally minerals) put together in a certain way. All other properties, such as color, density, and similar mass properties are only derived properties. Even structure is not entirely a primary property but is rather the reflection of changes—abrupt or gradual, horizontal or vertical—in texture and composition within one formation or between different formations. However, for practical purposes, structure can be considered as the third major property of sediments when describing entire sedimentary bodies.

A tenable (that is, objective and reproducible) classification of sediments must be based for hand specimens on composition and texture and for entire sedimentary bodies on composition, texture, and structure, with possible qualifications introduced by the addition of some of the subelements of texture (size, homogeneity) and possibly some of the principal derived properties, such as color.

There are altogether ninety-three important derived properties or parameters of sedimentary rocks, but most of these are quite unnecessary for purposes of megascopic identification.

## THE COMPOSITION OF SEDIMENTS

### MAJOR CONSTITUENTS

Just as it is impossible to understand the properties and probable behavior of an engineering structure without knowing of what materials it has been built, so it is equally impossible to understand a sedimentary rock without knowing what it is made of, and by this is meant knowing rather than guessing. Hence an adequate working knowledge of elementary mineralogy is an absolute prerequisite for the field study of sediments.

Although over one hundred and sixty different minerals have so far been identified in sediments, less than twenty mineral species form well over 99 per cent of the bulk of sedimentary rocks. It is rare indeed that more than five or six minerals occur in sizable amounts in any one rock. Thus the exact determination of the composition of a sediment (shales excepted) is not a particularly difficult affair.

The twenty minerals which form the bulk of the sedimentary rocks are known as "major" or "main" constituents. A major constituent is defined as one that forms at least 1 per cent of a rock. A further subdivision of major constituents into more abundant varieties, which form over 10 per cent of some common rocks, and less important ones, which form only 1–10 per cent, is shown in table 1.

In addition, approximately twenty other minerals, known as "accessory minerals," occur in very small amounts in sediments, although locally they may be of great importance. These accessory minerals, however, are difficult to study megascopically. A list of the major con-

stituents and of the more important accessories is given in table 1.

An understanding of sediments may be condensed to an understanding of the relationships which exist among the major constituents. Most of these relationships are established at the very beginning of the "life" of a sediment and generally are a function of its mode of formation.

Computations by the writer have shown that the average sediment consists of the following minerals and mineral families:

| | Per Cent |
|---|---|
| Quartz | 31.5 |
| Chalcedony (chert) | 9.0 |
| Feldspars | 7.5 |
| Micas and chlorite | 19.0 |
| Clay minerals | 7.5 |
| Carbonates | 20.0 |
| Iron oxides | 3.0 |
| All others | 3.0 |

These minerals are not distributed in sedimentary rocks in a haphazard manner but tend to occur in certain definite associations, which form a series of petrographic end-members.

TABLE 1*

THE COMMON MINERALS OF SEDIMENTS

| | Major Constituents | | Accessory Minerals (Less than 1 Per Cent of Rock) |
|---|---|---|---|
| | Over 10 Per Cent of Rock | Less than 10 Per Cent of Rock | |
| Detrital Minerals | QUARTZ<br>Microcline<br>CLAY MINERALS<br>(kaolin-bauxite)<br>FINE-GRAINED MICAS<br>(illite, sericite muscovite) | DETRITAL CHERT<br>Sodic plagioclase<br>(albite-oligoclase)<br>Coarse-grained micas:<br>muscovite<br>biotite<br>chlorite<br>Hematite<br>Limonite | "IRON ORES": MAGNETITE, ilmenite, DETRITAL LEUCOXENE<br>STABLE GROUP: ZIRCON TOURMALINE, rutile<br>UNSTABLE GROUP: APATITE, EPIDOTE GARNET, HORNBLENDE kyanite, sillimanite staurolite, titanite zoisite<br>MICAS: frequently occur as accessories rather than as major constituents |
| Chemical and Authigenic Minerals | CALCITE<br>DOLOMITE<br>ANKERITE | CHERT and opal<br>"SECONDARY" QUARTZ<br>GYPSUM and anhydrite, halite<br>Some hydromicas of the illite-sericite-chlorite series<br>Phosphates and glauconite, Siderite and some iron ores | ANATASE; authigenic rutile and leucoxene |

* Minerals printed in capitals are the relatively more common ones within each group.

PETROGRAPHIC END-MEMBERS

The composition of *all* sedimentary rocks can be reduced to two basic groups of end-members, which may be mixed in all proportions in a purely mechanical way.

1. A detrital fraction consisting of solid material brought in as *solid detritus* from outside the basin of deposition and precipitated through settling within this basin. In 99 per cent of the cases this detrital fraction is made up of silicates. The composition of this detrital fraction depends on the petrology of the source areas and the intensity (and effectiveness) of chemical weathering and erosion

within the source area, plus some modification during transport.

2. A chemical fraction existing as a *solution* within the basin of deposition and precipitated chemically. This chemical precipitation may proceed through inorganic or biogenic agencies. The chemically precipitated material may not move on the sea floor after its precipitation, and then it develops a crystalline texture. Or it may be shifted about by bottom currents (particularly true of organically precipitated rocks composed of formed from the consolidation of, let us say, a basaltic lava flow into which fell a shower of rhyolitic volcanic ash. The basalt would be the chemical end-member (from solution), the rhyolitic ash the detrital (solid) end-member. Such an igneous rock, although theoretically possible, would be a very rare freak indeed. But every aqueous sediment, without a single exception, has been formed exactly in this way, by a mixing of solid detrital and chemically precipitated material (fig. 1).

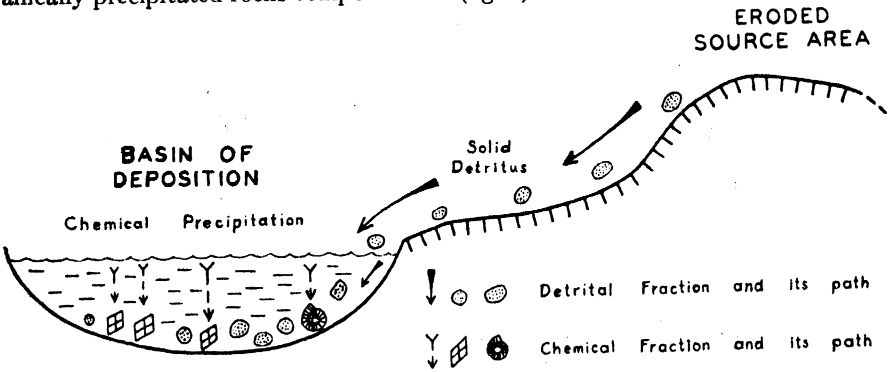



FIG. 1.—Any sediment is a mechanical mixture in every possible proportion of a detrital fraction, brought in *as a solid* into the basis of deposition from the source area, and a chemical fraction, precipitated *from solution* within the basin of deposition itself.

shell fragments), and then, although its origin is chemical, its texture will be clastic. More than three-quarters of this chemical fraction is made up of carbonates, and most of the balance of silica. Other constituents (glauconite, phosphates, iron oxides) may be very abundant locally but are relatively rare on a volumetric grand-total basis.

The fundamental difference between sediments and igneous rocks is that sediments consist of *two* major groups of end-members (detrital and chemical), whereas igneous rocks are made up of only one (generally chemical, tuffs excepted). The only possible direct analogy of a sediment within the igneous group would be a rock

DETRITAL AND CHEMICAL ROCKS

All sedimentary rocks are mixtures of detrital and chemical material. For megascopic purposes detrital material is defined as consisting of *clastic* silicates. The term "silicate" is used in the extended, so-called "modern," sense and includes both minerals of the free-silica group (quartz, etc.) and the combined silicates. If the detrital material exceeds 50 per cent, the rock is named a "detrital rock"; if it is less than 50 per cent, then the rock is a chemical rock (fig. 2). If the detrital material in a chemical rock fluctuates between 5 and 50 per cent, the chances are very great that this chemical rock has a clastic (or arenitic) texture. If the de-

trital material is less than 5 per cent, the chances are fair that the rock may have a crystalline texture, although a clastic texture is also very possible.

The proportions between chemical and detrital fractions may vary considerably over very short lateral or stratigraphic intervals, as shown in figure 3, which illustrates the simultaneous operation of two basic sedimentary processes and their results. The inclined line indicates the gradual decrease with distance (between points *1* and *8*) of the average grain size of the mechanically deposited detritus within a given basin of deposition. This average size drops from over 0.250 mm. (medium sand) at point *1* to less than 0.002 mm. (fine clay) at point *8*. The range of deposition of these clastic textural types is shown at the bottom of the diagram.

Vertical arrows show, on the other hand, the intensity of chemical precipitation of calcium carbonate at the same given points (*1*–*8*). The arrows indicate how much of the sediment at any given point consists of carbonate. For instance, at point *1* the average grain size of the detrital fraction is 0.3 mm., and the rock as a whole contains less than 2 per cent of calcium carbonate. The rock is a medium-grained sandstone. On the other hand, at point *7* the average grain size of the detrital fraction is less than 2 microns, and the rock contains over 85 per cent of carbonates; hence the rock here is an argillaceous limestone; at point *3* it is a sandy limestone; at point *5* a siltstone; etc.

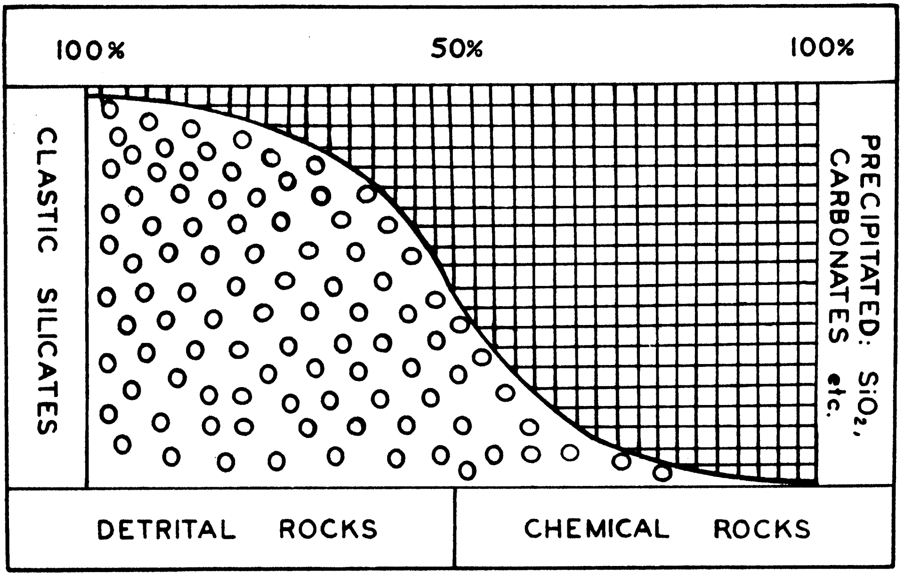



FIG. 2.—Relative amounts and character of volumetric changes between the detrital fraction (clastic silicates) and the chemical fraction (precipitated carbonates, $SiO_2$, etc.) in the detrital and chemical rocks, respectively.

This shows that the same two petrographic end-members, if mixed in variable proportions, may produce an almost infinite number of rapidly changing subtypes; each of these subtypes possesses

only a very local significance, and none is particularly helpful in classifying the basic sedimentary series of the region.

### THREE GROUPS OF DETRITAL ROCKS

As demonstrated elsewhere (Krynine, 1942, pp. 537–561; 1943; 1945, pp. 12–22), the detrital fraction (fig. 4) can be divided into three groups on the basis of composition. This division has been found to hold good in rocks of all ages, from all parts of the world.

1. Quartz, with or without detrital chert grains (liberated from eroded limestones).

2. Quartz plus detrital chert grains plus abundant rock fragments (generally micaceous low-rank metamorphic rocks, such as slates, phyllites, and schists and sometimes surface igneous rocks) plus an abundance of micas and chlorite, either as large flakes or, more commonly, as a micaceous or chloritic clay or clayey paste. Feldspar may be present.

3. Quartz plus a large amount of feldspar with a subordinate (20 per cent) amount of impurities (generally similar to the rock fragments described under 2) plus or minus some clay. In this case the clay is generally kaolinitic rather than micaceous (odor test).

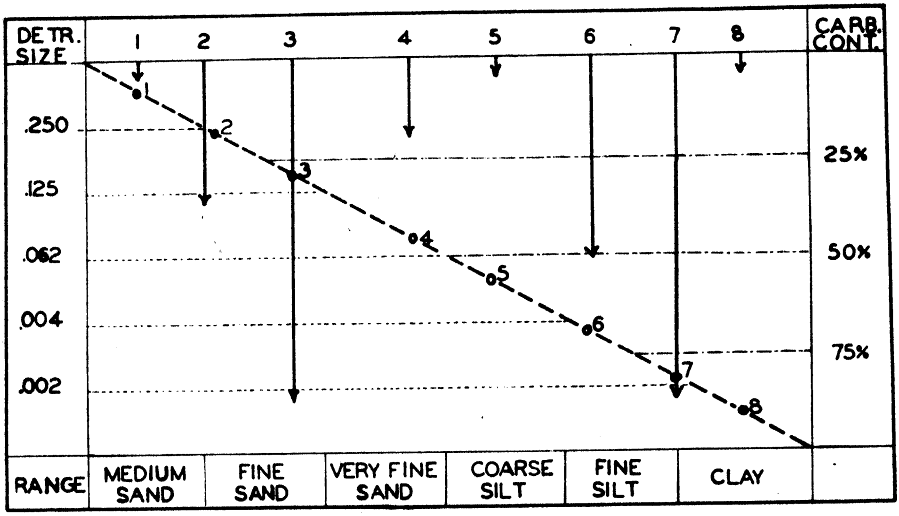



FIG. 3.—Local variations between chemical and detrital rocks over short lateral or stratigraphic intervals. *Left*, average diameter of detrital fraction in millimeters. *Right*, percentage of calcium carbonate in any given rock type (as shown by arrow). Rock types present are: (1) medium-grained sandstone (0.3 mm. av. grain size; 2 per cent of $CaCO_3$); (2) medium-grained calcareous sandstone (0.25 mm. av. grain size, 34 per cent $CaCO_3$); (3) sandy limestone (85 per cent of $CaCO_3$; 0.13 mm. av. size of detrital particles in it); (4) very fine calcareous sandstone (0.065 mm. av. size; 20 per cent of $CaCO_3$); (5) sandy siltstone (0.05 mm. av. size; 2 per cent $CaCO_3$); (6) marl (50 per cent of $CaCO_3$; 0.035 mm. av. size of detrital particles in it); (7) argillaceous limestone (85 per cent of $CaCO_3$; 0.002 mm. av. size of detrital particles in it); and (8) fine shale (0.001 mm. av. grain size; 2 per cent of $CaCO_3$).

These three mineral assemblages are used as the basic classification scheme and correspond, respectively, to the quartzite, graywacke, and arkose series of sediments as described in part 2.

## THE TEXTURE OF SEDIMENTS

### DEFINITION

Texture is the interrelationship of the individual particles of a rock. Texture or

"grain" can be defined as the fabric pattern of an aggregate of mineral particles.

Texture depends partly upon composition (i.e., upon types of particles and their inherent properties) and partly upon the mutual arrangement of the particles. Texture may be either clastic or crystalline.

1. *Clastic particles* or grains, either of detrital or of chemical origin, have been rolled around and moved to and fro before coming to rest within the sediment. They are always somewhat abraded or may be entirely broken up into fragments.

2. *Authigenic or crystalline (granular) particles* have been directly precipitated from solution on the basin bottom and have remained there or otherwise have grown within the sediment and thus have not substantially been moved after deposition and consolidation of the rock.

All detrital rocks and all impure (sandy) chemical rocks are clastic. About 80 per cent of the pure chemical rocks show a clastic texture; the balance have a crystalline texture.

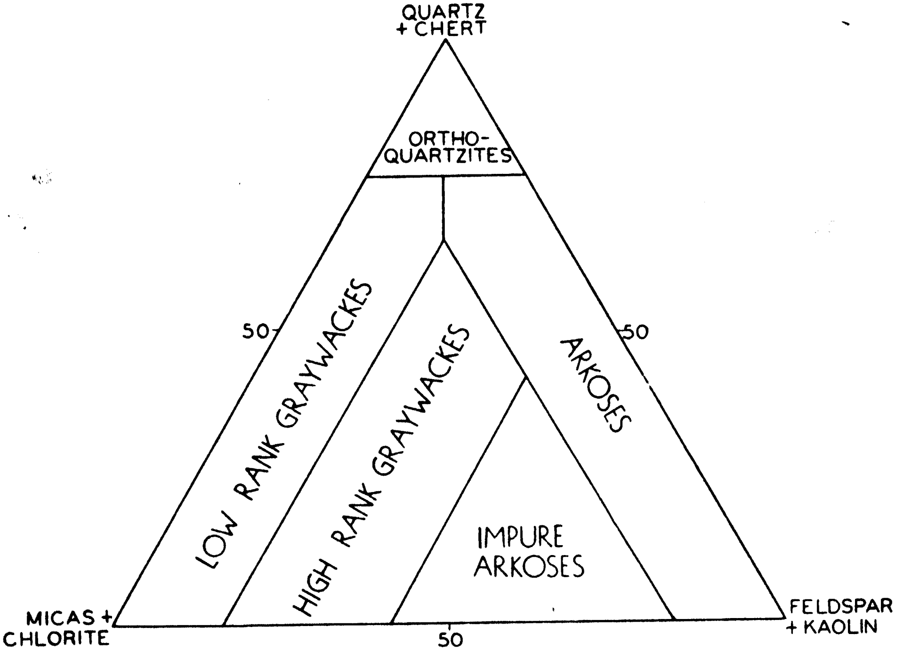



FIG. 4.—Mineral composition of the detrital fraction of the major petrographic series of sedimentary rocks. Note that the micas and chlorite may occur as rock fragments, as large, loose, individual flakes, or as a micaceous clayey paste.

TEXTURAL ELEMENTS

The mineral constituents of a rock can be grouped into three main textural elements which build up the rock and give it its appearance. This grouping is on the basis of the *relative* (not the absolute) size and the relative position in space of the constituents.

1. *Grains.*—Grains or individual particles are the basic units of texture. If particles are of different sizes, occurring

in distant orders of magnitude, the term "grains" is restricted to the *larger* particles.

2. *Matrix.*—If particles are of different orders of magnitude, then the term "matrix" is used for the smaller individual units which fill the interstices between the larger grains. Hence "matrix" is a relative term only, and "matrix" by itself does not exist without the simultaneous presence of larger grains. If the particles are very small but of the same size, then they are to be termed "grains," not "matrix."

3. *Cement.*—Cement is the authigenic chemical precipitate which infiltrates around grains and matrix. In clastic rocks (which form well over 95 per cent of all sediments) grains and matrix are always clastic (either of detrital or of chemical origin), and cement is always chemically authigenic. In crystalline sediments the periods of formation of grains, matrix, and cement may overlap, although frequently the distinction of these separate generations is possible. An almost direct analogy can be worked out between the formation and morphology of crystalline sedimentary rocks and that of igneous porphyritic rocks.

All these terms are relative, and the term "grain" has no absolute size connotation.

In coarse- and medium-grained clastic rocks the possible combinations are: (1) grains (or pebbles), matrix, and cement; (2) grains and matrix; (3) grains and cement (fig. 5).

In fine-grained clastic rocks the possible combinations are: (4) grains alone (not matrix and cement or matrix alone, as frequently misunderstood by beginners); (5) grains and cement.

In crystalline rocks the possibilities are: (6) grains alone; (7) grains and cement; (8) grains and matrix; (9) grains, matrix, and cement.

### CLASTIC TEXTURE AND ITS COMPONENTS

The clastic silicate fraction in a detrital rock may be loose (unconsolidated) or consolidated either by simple adhesion between the more plastic, finer-grained constituents of the matrix or by introduction of a foreign chemical cement. When this cement exceeds 50 per cent, the rock becomes a chemical rock.

Hence the two groups of detrital rocks are these: (*a*) without chemical cement (loose or consolidated only by adhesion or reorganization of the matrix) and (*b*) those consolidated through the action of chemical cements.

The two principal types of matrix are the argillaceous (kaolinitic-bauxitic) type and the micaceous-chloritic type. Both may be heavily loaded with very fine particles of quartz. A schematic representation of these two types of matrix is given in figures 13 and 14.

Both are "clayey" in the purely physical sense of the word, i.e., exceedingly fine grained. This matrix may be colored black or green by carbonaceous matter, biotite, or chlorite, or red and brown by ferric oxides. Because in this case the iron oxide acts only as a pigment and not as a cement, the term "red" rather than "ferruginous" is to be used (many black rocks contain much more iron than do some red ones).

There are two principal types of chemical cement: silica and carbonate. Their typical occurrence and relationships are shown in figure 12.

The silica cement may occur either as quartz overgrowths on quarts grains, thus greatly increasing the apparent angularity of a sediment, as chalcedony (chert), or as opal.

The carbonate cement is calcite or dolomite or, much more frequently than is generally expected, an iron carbonate of the ankerite or siderite class. Calcite is distinguished from the other carbonates by the dilute-acid test, whereas iron carbonates generally show conspicuously oxidized surfaces.

In addition, a third group of miscellaneous chemical cements includes all other chemically precipitated material. Although gypsum, anhydrite, and even halite cements may be well displayed locally, in general only glauconitic, phosphatic, and ferruginous materials are important. By "ferruginous" cement is meant clay-free hematitic or limonitic cement, which acts as an actual bonding agent and not only as a red pigment for a clayey matrix. In many cases this can be determined without too much difficulty with a hand lens.

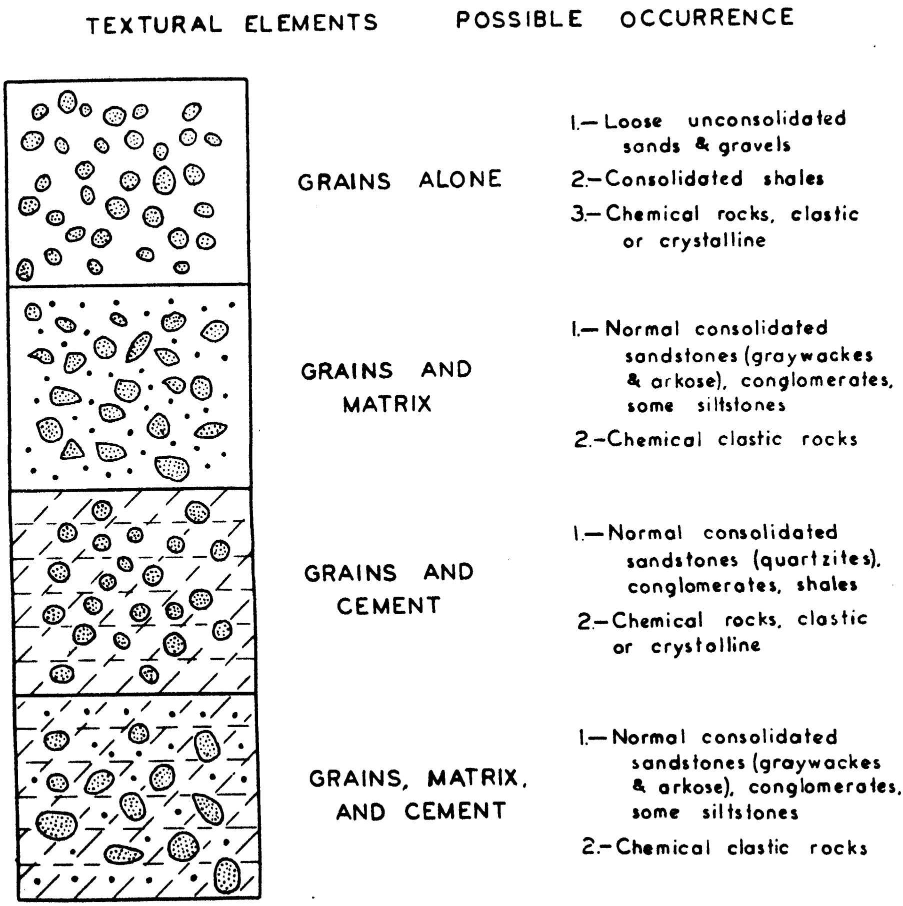



FIG. 5.—Textural elements and their possible combinations and occurrences

TEXTURE OF DETRITAL ROCKS

*Grain size.*—Detrital rocks occur in three grade sizes: (*a*) coarse grained or conglomeratic with an average size of over 2 mm., which in practice means grains of a much larger diameter in order to average up to 2 mm. with the finer matrix; (*b*) medium grained or sandy (from 2.0 to 0.0625 mm.); (*c*) fine grained or silty or clayey (average below 0.0625 mm.).

Since these textural changes occur gradually, intermediate stages can be set. These subtextures, as shown in table 3, are based on relative sizes and on the introduction into a grade size of elements from the preceding or following grade sizes. A graphic representation of the quantitative textural data of table 3 is given in figure 6.

In the conglomerate (psephite) class, common conglomerates are defined as being between 4 and 64 mm. in diameter, boulder conglomerates over 64 mm., and fine conglomerates below 4 mm. The introduction of over 20 per cent of sand, silt, or clay makes the rock a sandy, silty, or clayey conglomerate (the so-called "boulder clays," tills, and many fanglomerates belong in these categories).

Sandstones (psammites) are not easily subdivided megascopically on the basis of size, although the terms "coarse" or "fine" are easily applicable to the end-members of the sandy series (above 1 or 1½ mm. and below ½ mm. in diameter, respectively).

A more refined division into very coarse, coarse, medium, fine, and very fine sand is possible if the specimen under study is compared directly with a series of "yardsticks" or sand samples of known grain size, properly mounted in glass vials.

In defining grain size and especially the average grain size of a sandstone, it must be remembered that in a hand-lens or even binocular-microscope study the finer (clayey) fractions are very difficult to interpret, and hence the average grain size should be based on the size of the sandy fraction above, thus producing an average size which will always be somewhat coarser than the median diameter of a conventional mechanical analysis. These two figures should not be confused but clearly differentiated in descriptions.

In addition, depending on the type and amount of admixture, several subtextures may be established for sandstones. A rock may be termed a "normal," "conglomeratic" (over 20 per cent of pebbles), "pebbly" (over 10 per cent of pebbles), "silty" (over 20 per cent silt), or "clayey" sandstone (over 20 per cent clay).

In the finer clastics (pelites) the siltstones are described megascopically as gritty, and the shales and clays as smooth and unctuous to the touch. Sandy siltstones contain over 20 per cent of sand. Sandy shales on the basis of the preceding definition are impossible (since such "shales" would be gritty and hence automatically be called "siltstones"). Silty shales (semigritty or not perfectly unctuous), however, do exist.

Siltstones occur in two definite fabric patterns: (*a*) siltstones proper, consisting of relatively well-sorted silt particles, and (*b*) microconglomerates (or rather microbreccias), made up of relatively coarse sand grains in a very fine silty or clayey matrix. These two types, however, would be difficult to distinguish megascopically but for the fact that siltstones proper are generally somewhat more micaceous than are microbreccias.

*Grain shape.*—Coarse clastics can be divided on the basis of their angularity into conglomerates proper (rounded) and

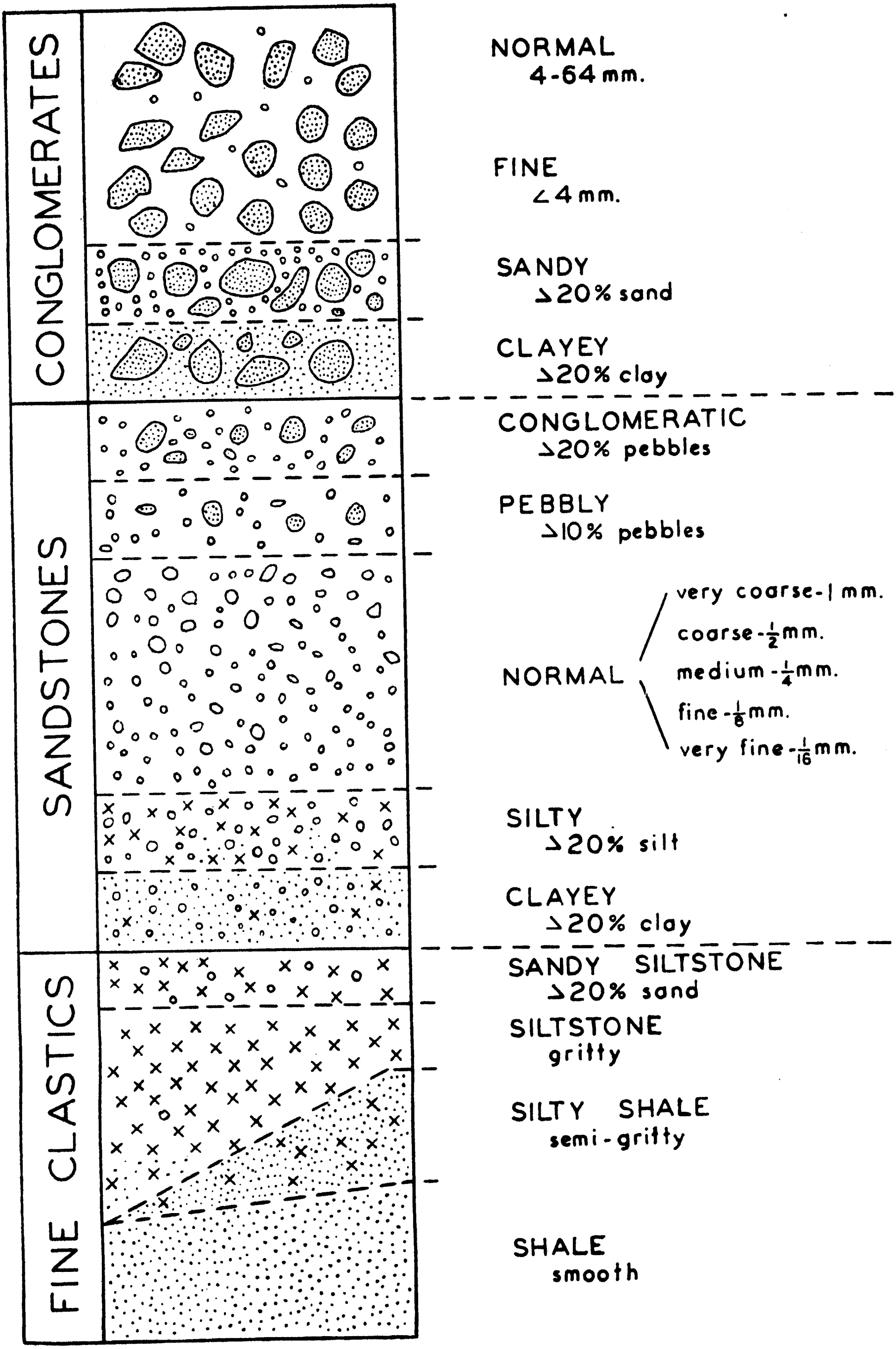



FIG. 6.—A megascopic textural division of clastic rocks, according to grade size

breccias (angular). In the medium-grained clastics this division may be somewhat difficult because many so-called "grits" are produced by overgrowths of secondary silica. Hence the terms "rounded" or "angular" should be used for sands rather than such specific names as "grit." All fine-grained clastics are angular, but this fact cannot be easily observed with a hand lens.

With a little effort it is possible to describe most sands as angular, subangular, subrounded, or rounded. This can be accomplished either through the use of appropriate comparison specimens or sim-

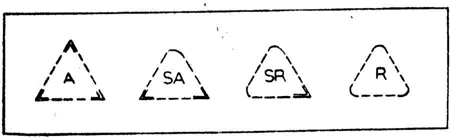



FIG. 7.—Definitions of roundness in clastic particles: *A*, angular: all edges are sharp; roundness coefficient is 0. *SA*, subangular: one-third of edges are smooth; roundness coefficient is 33. *SR*, subrounded: two-thirds of edges are smooth; roundness coefficient is 66. *R*, rounded: all edges are smooth; roundness coefficient is 100.

ply on the basis of the following arbitrary definitions, as shown in figure 7:

Angular—all edges of a grain are sharp.
Subangular—one-third of the edges are smooth.
Subrounded—two-thirds of the edges are smooth.
Rounded—all edges are smooth.

By assigning the values of 0, 33, 66, and 100 per cent to these respective terms and by multiplying these values by the number of grains of each class within one specimen, it is possible to describe the roundness of a sandstone in quantitative numerical terms, even on the megascopic or at least binocular-microscope level. This procedure is perfectly feasible for descriptions of well logs.

Roundness should not be confused with sphericity. At the megascopic level the sphericity of a grain can be defined as follows:

Equant—length of grain is less than $1\frac{1}{2}$ times its width.
Elongated—length $1\frac{1}{2}$ to 3 times its width (prismatic) or thickness (tabular).
Acicular or platy—length is over 3 times width or thickness, respectively.

Sphericity (i.e., the ratios between length, width, and thickness) is very difficult to evaluate correctly in particles of sand size or finer, since there may be considerable differences between the appearance of a solid particle when viewed in three or only two dimensions. Hence sphericity terms should be used with discretion at the megascopic level.

### TEXTURE OF SANDY CHEMICAL ROCKS

In the chemical rocks clastic texture always occurs in the sandy or arenitic types (containing from 5 to 50 per cent of clastic silicates). The grade sizes used (coarse, medium, or fine) are the same as in the detrital rocks. If necessary, subtextural terms of the same kinds as those used for the detrital rock should be introduced for the sake of precision. The crystalline constituents should be described separately.

### TEXTURE OF PURE CHEMICAL ROCKS

In the pure chemical rock (less than 5 per cent of clastic silicates) the texture may be either clastic or crystalline. If the rock contains more than 10 per cent of fragmental particles of any kind (fossil shells, intra-formational pebbles, oölites, etc.), then very probably the texture is clastic, and the same terminology is used as before.

However, the large amount of crystalline cement present should be carefully described according to the terminology suggested in the next paragraph. The rock is then referred to as a "clastic limestone," "clastic dolomite," etc.

If the texture is definitely crystalline, then the rock is no different texturally from an igneous rock. To avoid confusion, however, the terms "phaneritic" and "aphanitic" should not be used. Instead, the terms "coarsely crystalline" (over 4 mm. in diam.), "medium crystalline" (1–4 mm.), and "finely crystalline" (below 1 mm.) are suggested.

Furthermore, the degree of homogeneity of the size distribution of the crystalline elements (granules) should be expressed by the terms "even" or "uneven" (or heterogeneous), with adequate quantitative notation as to the amount of crystalline particles in each major grade size. The uneven or heterogeneous texture of crystalline sediments corresponds to the porphyritic texture of igneous rocks.

Finally, the shape of the crystalline particles should be noted, primarily their degree of sphericity (as suggested above), and their idiomorphism.

A graphic presentation showing differences between clastic and crystalline textures in chemical rocks is given in figure 15.

## THE COLOR OF SEDIMENTS

The color of a sediment is the most obvious, most striking, and most easily observable property. A purely objective approach capable of yielding reproducible results is possible only on the basis of a color dictionary or at least of simplified color charts. A very fine study of color in rocks, discussing all previous work and giving a complete list of references and suggestions, has recently been made by De Ford (1944, pp. 128–137) and should be read by all geologists.

An extremely simplified summary of color in rocks follows.

1. Any color is a mixture of two basic elements: a pigment and a neutral background, which dilutes the pigment. The proportion of pigment to neutral background is known as "saturation," "purity," or "chroma."

2. A pigment is a mixture of several colors, principally blue, green, yellow, and red. The hue depends on the proportions of these four basic colors.

3. The neutral background is a mixture of white and black, in practice, therefore, being finally some shade of gray. The relative proportions of light and black will determine the lightness (or brilliance or value) of the color.

These items are usually plotted on a triaxial ellipsoid, with a vertical axis representing the white and black poles of the neutral background cutting across a central circular cross section, along the periphery of which are disposed the different hues.

This scheme can be greatly simplified by showing these relationships in two dimensions only, as sectors of a circle on a so-called "pie-shaped" diagram, as shown in figure 8.

Figure 8, *A*, shows the purity (saturation or chromatic value) of a color. In the left portion the color will be vivid or deep (high porportion of pigment); in the right portion the color will be drab or grayish (low proportion of pigment to background).

Figure 8, *B*, illustrates the difference between a light and a dark color as conditioned by the change in proportion of black to white within the neutral background.

Figure 8, *C*, shows the formation of different hues by mixing different amounts of red and yellow within the pigment.

Figure 8, *D*, reveals the combined effect of all these factors by showing the difference between orange (in which pigment predominates over background

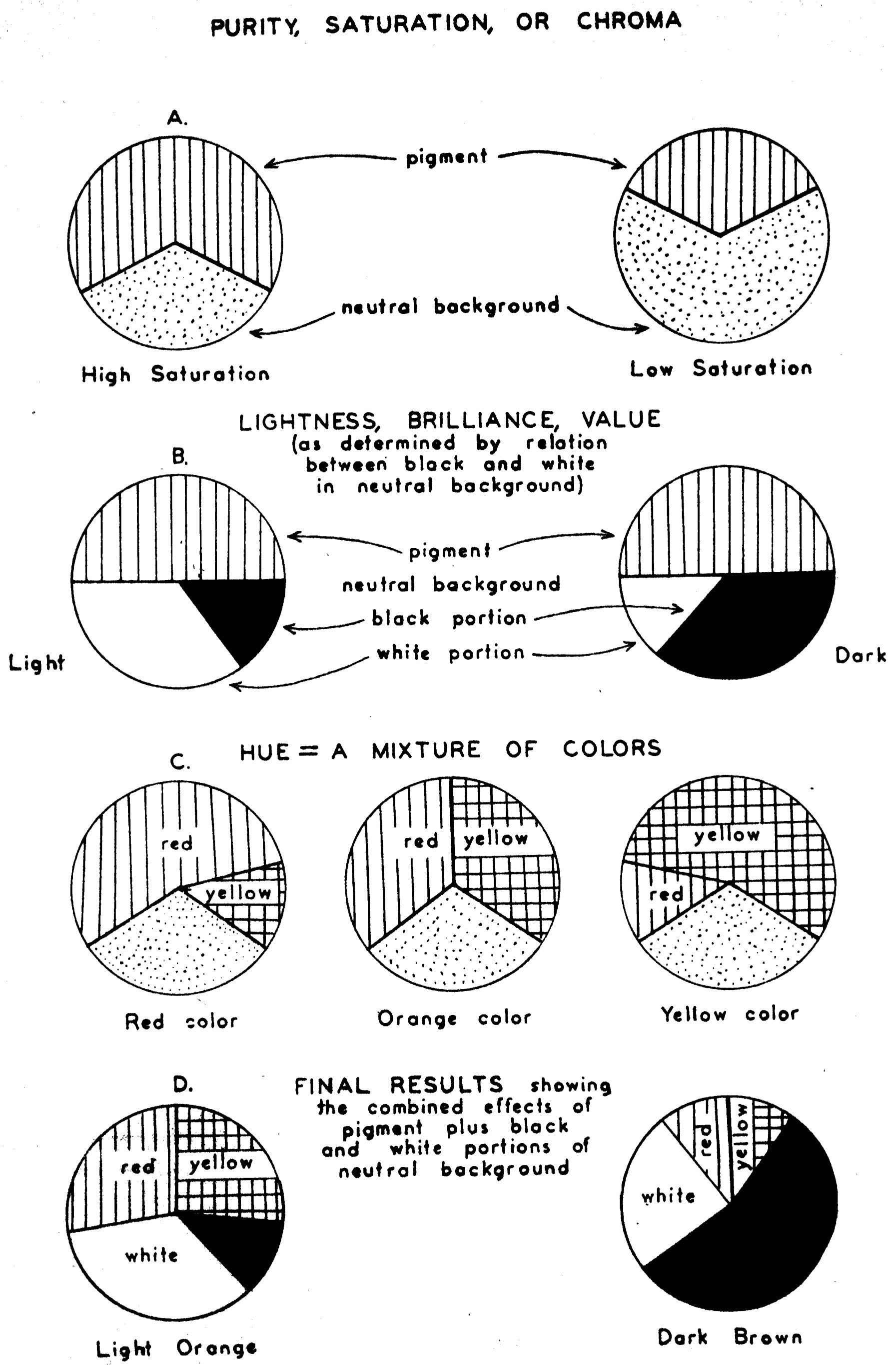



FIG. 8.—A two-dimensional interpretation of color showing: *A*, purity (ratio of pigment to background). *B*, brilliance (ratio of white to black in background). *C*, hue (ratios of basic colors in pigment). *D*, final results of interplay between *A*, *B*, and *C*.

and white over black within the background) and brown, in which the proportions of background to pigment and of black to white within the background have been reversed.

An understanding of these simple fundamentals of color is necessary to enable one to describe and compare intelligently various colors in rocks.

For rapid megascopic examination of a preliminary nature a simple approach to the description of color in rocks is possible on the basis of four basic colors and eight intermediate hues plus two neutral shades (light and dark).

The following color scheme has been found to be applicable without undue confusion or loss of time in the field:

1. The basic hue sequence consists of blue, green, yellow, and red.

2. The neutral shade sequence is light and dark, which when taken alone produced the sequence

white—gray—black.

When applied to pigments this sequence will be

light color—drab color—dark color.

3. The intermediate hues are

Between blue and green—greenish blue and bluish green
Between green and yellow—yellowish green and greenish yellow
Between yellow and red—orange
Between red and green—purple
Between red and blue—lavender and purple

These colors and hues are based not only on a study of color in the abstract but, to some extent, on the mineralogical basis of color. For instance, purple is a typical color produced by a mixture of green and red minerals (such as hematite and chlorite or hematite and glauconite) and can be used as a criterion for stratigraphic and genetic purposes.

Each of these four basic and five intermediate hues can be qualified by the terms "light" and "dark" (no qualification for the normal shade).

An additional terminology is necessary when dealing with brown and red rocks, even at the megascopic level. The normal color sequence in this case is red, orange, yellow. If this pigment is highly diluted by a neutral background, in which the black portion predominates, the following sequence of three additional colors is arrived at: cherry red, brown (or drab, depending upon relative lightness), olive.

If these fundamentals are kept in mind, then color can be used as a powerful tool in describing rocks because the variations in "shade" reflect variations in mineral composition. It must be kept in mind that a finer-grained rock will always show a more vivid color than will a coarser-grained one because of the better distribution of the pigment throughout the rock.

For more precise work the standards used in soil and rock nomenclature as reviewed and defined by De Ford should be applied.

Color should always refer to a dry fresh surface with a degree of roughness or smoothness characteristic of the rock. Wet, polished, weathered, chipped, or other non-natural surfaces give abnormal colors and should always be identified as such if used for color determination.

## THE STRUCTURE OF SEDIMENTS

### DEFINITION

*External and internal morphology.*—Structure in sediments may be defined as the visible expression of nonuniformity in texture and composition and hence is, strictly speaking, a second-order property derived from these fundamental two. However, since structure is a pri-

mary and fundamental attribute of large sedimentary bodies or rock masses, it can be considered as the *third basic property* of sediments.

Different types of structure, i.e., structural elements, or "structures" for short, can be classified as follows: (1) *morphologically* into external and internal or (2) *genetically* into primary and secondary.

The *external morphology* of a sedimentary body can be defined by its extent (size), its shape, and the character of its boundaries. Since the size and shape of a sediment are directly related to its composition and texture and since these properties can be tangibly expressed in quantitative terms, a quantitative classification of sedimentary bodies based on external morphology is also proposed. This classification is for large-scale field study and obviously does not apply to hand specimens.

### SIZE OF SEDIMENTARY BODIES

The data expressing the size of a sedimentary body include the horizontal *extent* (length, width, area), *thickness*, and *volume*. Quantitative definitions of these terms are given in table 2.

In stratigraphic work very frequently these size terms (large or thick, small or thin) are used in a purely subjective way, depending upon the relationships existing within a given area or the training and idiosyncrasies of the worker. A "thick" formation within the limestone series of the Mid-Continent Paleozoic may become a very "thin" one in the arkoses of the California Tertiary. Hence objective, reproducible, and understandable terms like those of table 2 become necessary.

### SHAPE OF SEDIMENTARY BODIES

The shape of a body of deposition can be expressed either in terms of configuration or through the relationships existing between the different size factors.

A. The external *configuration* of the body is described by such terms as "lens," "cone," "fan," "delta," and various types of prisms (from bars and channels to so-called "blanket" and "sheet" deposits). The meanings of these purely descriptive terms have been discussed and established adequately in textbooks on geomorphology and sedimentation, and the student is referred to these.

B. The *relation between area and volume*, another fundamental shape factor of a sedimentary body, can be expressed best as the *ratio between width* (i.e., extent measured across the strike) *and thickness*. On this basis, regardless of curvature or external configuration, there are four types of sedimentary bodies (fig. 9):

1. *Blankets*, when the ratio of width to thickness is *over* 1,000 to 1 (and up to 50,000 to 1). An example is the Lower Devonian Oriskany sand from the Appalachian region, which is over 400 miles long at the outcrop, has a width of at least 100 miles and a thickness of less than 50 feet, or a ratio of width to thickness of 10,000 to 1.

2. *Tabular bodies*, when the ratio of width to thickness is between 50 and 1,000 to 1. Examples of this are practically all the individual formations and sandy bodies in geosynclinal deposits (not their aggregates which may reach a very large thickness).

3. *Prisms*, when the ratio of width to thickness is *less* than 50 to 1 and may go as low as 5 to 1. Examples of this are most arkoses that are found in fault blocks, such as the Triassic of Connecticut (width 25 miles or less; average thickness 8,000 feet up to 16,000 feet; or a ratio of 15 to 1 or less). Prismatic deposits are typical of orogenic sediments

formed during periods of intense structural deformation of the earth's crust.

4. *Shoestrings*, when the ratio of width to thickness is less than 5 to 1, usually of the order of 1 to 1 or even less, because the thickness may exceed the width. Examples of this are most channel, many bar types of deposits, and fanglomerate belts, found within the larger tabular or prismatic bodies. The more extensive and flatter sandy bodies are generally produced by the coalescence of smaller and thicker ones. For instance, marine blanket sands represent the horizontal welding of many parallel prismatic shore lines during a long, continuous period of overlap or off-lap. Glacial deposits of the ground moraine type and ash beds are about the only exception to this rule. However, in chemical rocks, very widespread initial bodies of the blanket type are possible.

TABLE 2

A CLASSIFICATION OF SEDIMENTARY BODIES ACCORDING TO SIZE

| Definition | Length (Along Strike) (Miles) | Width (Across Strike) (Miles) | Area (Sq. Miles) | Thickness (Feet) | Volume (Cubic Miles) |
|---|---|---|---|---|---|
| Large (or thick)....... | >100 | >50 | >10,000 | >500 | >500 |
| Medium.............. | 20–100 | 5–50 | 100–10,000 | 100–500 | 1–500 |
| Small (or thin)........ | < 20 | < 5 | <100 | <100 | <1 |

#### MISCELLANEOUS STRUCTURAL PROPERTIES

Other elements of external structural morphology of sediments comprise boundaries: conformable or concordant (real or apparent), unconformable or discordant; lensing, pinching out, etc.

Elements of internal structural morphology in sediments (or "structures" for short) include layering or stratification (which should be described in terms of thickness and pattern); surface features along deposition planes (ripple marks, mud cracks, etc.); inclusions and internal growths (clay galls, pisolites, concretions, fossils); internal breaks (channeling, slumping features, intra-formational conglomerates, solution zones, surfaces of weathering, shrinkage, and expansion features); zones of abnormal concentration of any one of the preceding items or of such properties as sorting or porosity; structural attitude; and so on. These structural elements warrant detailed study; but, since most of these features are of no particular help in classifying or naming a sediment, they are omitted from the present discussion, and the reader is again referred to the standard text and reference books on this subject.

## PART 2. THE THREE MAJOR SERIES OF SEDIMENTARY ROCKS

Things that have a common quality ever quickly seek their kind.—MARCUS AURELIUS.

#### BASIS OF CLASSIFICATION

The purpose of the present study is to introduce a purely descriptive classification of sedimentary rocks suitable for objective megascopic and field use. It so happens, however, that this classification is also genetic, since every major sedimentary series is related to a definite degree of intensity of deformation of the earth's crust (diastrophism), operating during the deposition of this rock series.

Since, however, the formulation and proof of these genetic concepts involves the presentation of a considerable body of detailed microscopic and paleogeo-

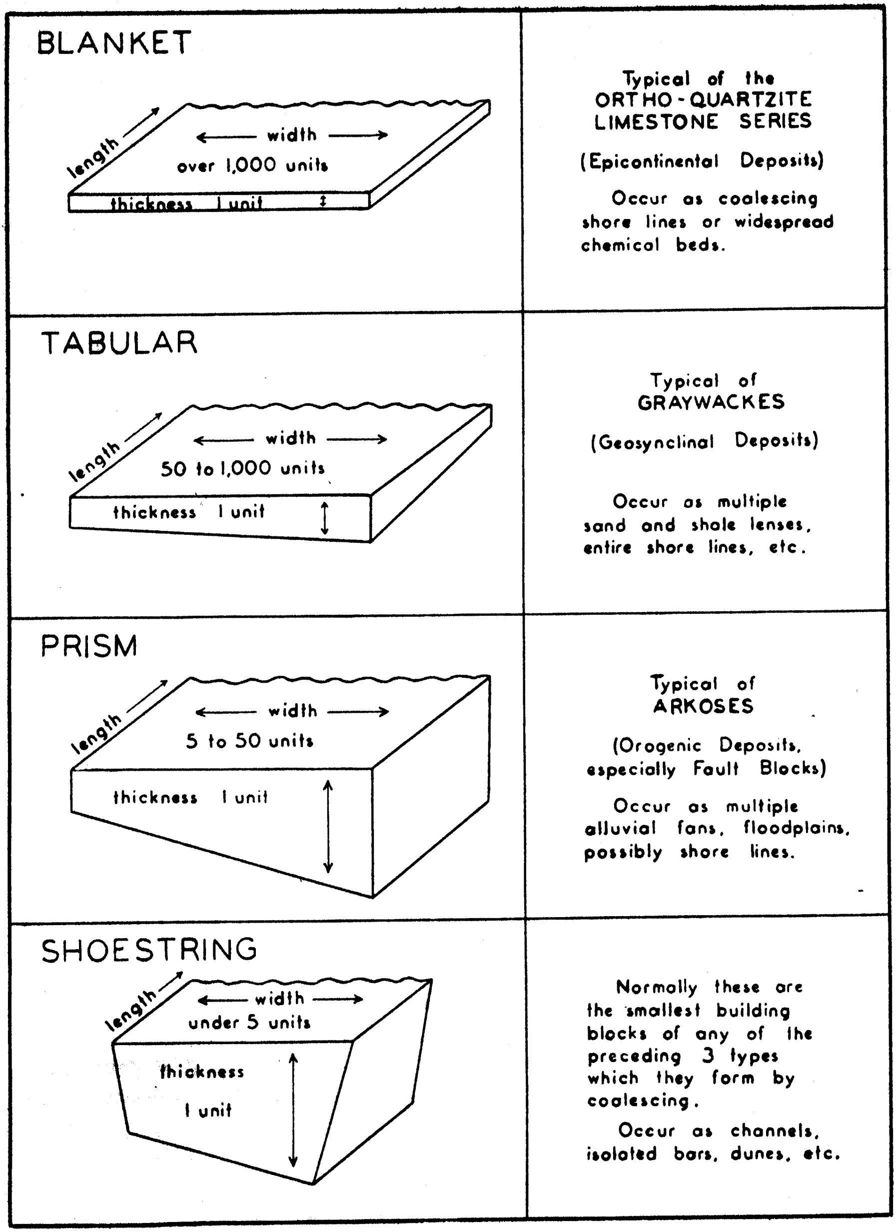



FIG. 9.—External morphology of sedimentary bodies, showing scheme of simplified geometric relationships between blanket, tabular body, prism, and shoestring.

graphic evidence, it has been thought best to refrain from any fundamental discussion on this subject in the present paper. Hence the material in part 2 is designed only as a help for an easier understanding and assimilation of the classification table (table 3). The present article presents a purely descriptive megascopic classification scheme and not genetic concepts. Some of the basic ideas underlying the explanation of these genetic relationships have been discussed in a preliminary way elsewhere (Krynine, 1942, pp. 537–561; 1943; 1945, pp. 12–22) and are going to be presented in considerable detail in a forthcoming publication.

The present study aims at a classification of sedimentary rocks sufficiently detailed for an understanding of the relationships among sediments observable with a hand lens. This classification is essentially founded on mineral composition (reinforced by texture) and is presented in table 3. It is based, first, on the preliminary division of sediments into two main classes—detrital and chemical—and, second, on the fundamental division of the detrital fraction of either class into three main types of mineral assemblage which will correspond to the three major sedimentary series:

1. The orthoquartzite series, made up of quartz with or without detrital chert grains. Most chemical rocks also belong to this series, which can be expanded thus into a quartzite-limestone series.

2. The graywacke series, made up of quartz plus chert grains plus abundant rock fragments (generally of low-rank metamorphic rocks, such as slates, phyllites, and schists and sometimes surface igneous and volcanic rocks) plus an abundance of micas and chlorite either as large flakes or, more commonly, as a micaceous or chloritic clay. Feldspar may or may not be present.

3. The arkose series, made up of quartz plus large amounts of feldspar with a subordinate (20 per cent) amount of impurities, generally similar to the material described under 2. The clay in the arkoses is generally kaolinitic rather than micaceous or chloritic, this being readily determined by an odor test.

A diagrammatic representation of the relations existing among the major mineral constituents within the sandstone member of these main sedimentary series is given in figure 10.

Figure 11 shows the total mineral composition of four typical representatives of the sandstone members of these classes. The data of Figure 11 are given without differentiating between the detrital and the chemical components within any of the mineral groups.

### QUARTZITE SERIES

The quartzite series of sediments can be popularly defined as being made up of "clean sands." It contains one detrital (clastic) end-member, consisting almost exclusively of quartz, occurring generally as medium-sized grains. These quartz grains (fig. 12) are cemented by at least one chemical end-member (cement), generally secondary silica or a carbonate (dolomite or calcite). The quartz grains are well rounded and excellently sorted.

The term "quartzite" refers to an orthoquartzite or primary sedimentary quartzite, as contrasted with "metaquartzite," which is a metamorphic rock. The vast bulk of quartzites in the earth's crust are orthoquartzites rather than metaquartzites.

Orthoquartzites, regardless of the crystallinity of their cementing material are normal sediments and have no more metamorphic significance than does a deposit of rock salt or gypsum (which are also highly crystalline rocks). The er-

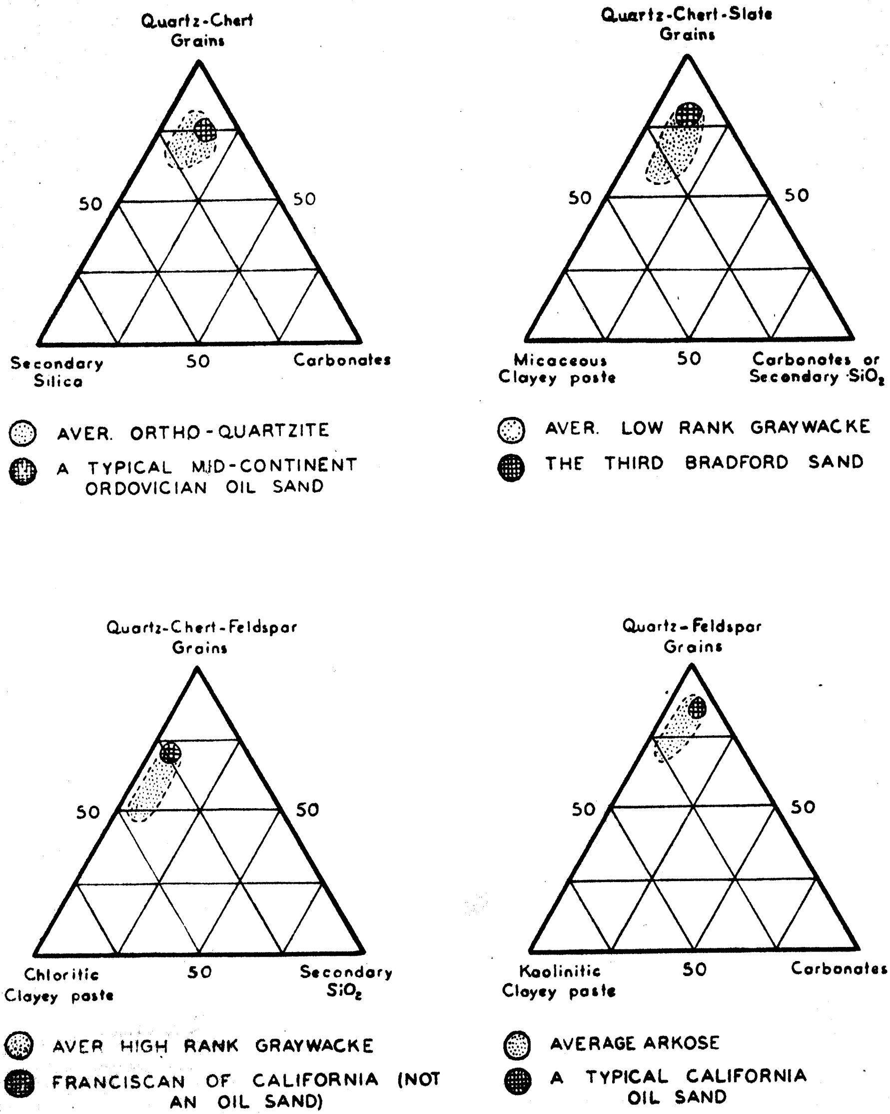



FIG. 10.—Mineral composition in terms of petrographic end-members for average sandstones and for some typical oil sands related to these sandstone classes, showing their respective fields of occurrence plotted on ternary diagrams for the orthoquartzite, graywacke, and arkosic series.

roneous and, unfortunately, fairly widely held misconception that most—if not all—quartzites are metamorphic rocks has adversely affected geologic judgment in evaluating the petroliferous possibilities of many a sedimentary province.

A so-called "textbook" quartzite in the restricted sense is a rock completely and solidly cemented by secondary quartz, so that the rock breaks across the grain (since the cementing matter is just as tough as the grains themselves).

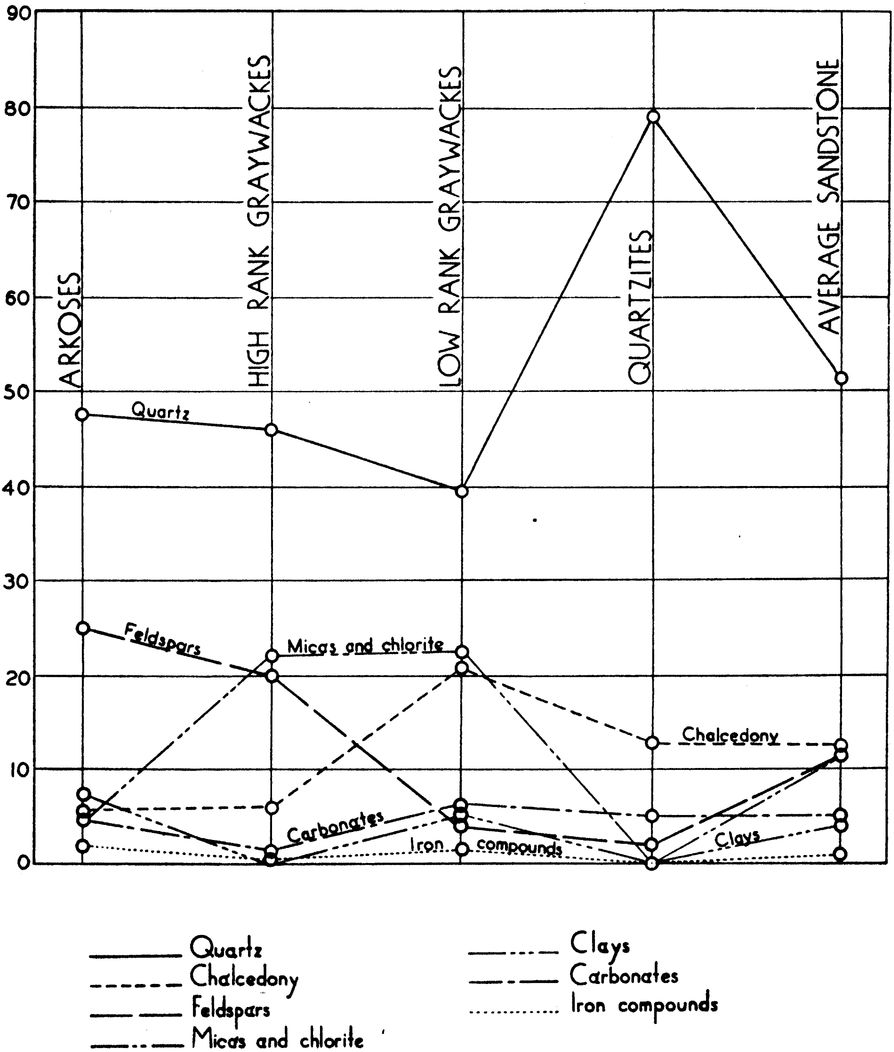



FIG. 11.—Average mineral composition of sandstones in the three major sedimentary series. Percentages shown refer to *total* composition and include both detrital and chemical constituents.

However, even the hardest and most siliceous quartzites contain some carbonate cement as an additional petrographic end-member, and, as the proportion of this nonsiliceous cementing material increases, the orthoquartzites begin to show more diversification in composition and less cohesiveness.

The fabric pattern of the average *oil-bearing* quartzite or quartzitic sandstone is that of rounded quartz grains rigidly but incompletely cemented by secondary $SiO_2$ (there is no porosity in a pure quartzite, 100 per cent cemented by $SiO_2$) with some additional parts of the pores filled by carbonates.

Structurally, the quartzitic sands occur most commonly in thin but extensive blankets. They are generally related to the quiescence (peneplanation and near-peneplanation) stage of diastrophism. The average grain size of an orthoquartzite is between 0.5 and 0.125 mm.

GRAYWACKE SERIES

The graywacke series of sediments can be popularly defined as a "dark dirty sand" or as a "pepper-and-salt sand" for the lighter-colored but strongly cherty varieties.

Graywackes (fig. 13) contain two or more major detrital clastic end-members: grains of quartz, chert, slate, schists, phyllites, etc., and a fine-grained matrix made of finely divided particles of micas (illite, sericite, muscovite, biotite, chlorite). Chemical end-members (cements) are subordinate and may be absent altogether. Feldspar is almost absent in the common or low-rank graywackes but may be abundant in the high-rank graywackes. In this case it is usually a sodic plagioclase.

The grains of a graywacke are medium to fine, generally very angular, and frequently elongated, and the sorting and sizing are very poor. The average grain

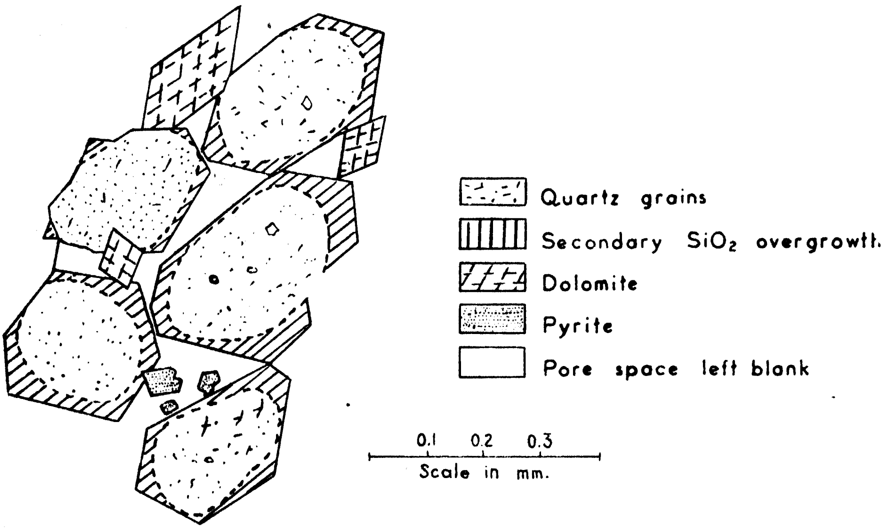



FIG. 12.—Schematic representation of an orthoquartzite (magnification 80×) showing development of pseudo-angularity through the appearance of secondary quartz overgrowths around originally rounded quartz grains and also the beginning of pitting and etching (at times mistaken for eolian frosting) on quartz grains through incipient replacement by dolomite or pyrite.

size of sandstones of the graywacke class fluctuates between 0.25 and 0.06 mm.

The fabric pattern of an average graywacke is a hodgepodge of angular, elongated small pebbles and rock fragments, quartz and chert grains, heavily loaded with mica flakes and bonded together by much so-called "clay," which, in reality, consists mostly of finely chopped-up or recrystallized micas or chlorite.

spathic graywackes are formed in narrow and rapidly subsiding geosynclines, frequently connected with volcanic activity (eugeosynclines). The graywackes are by far the most important kind of sandstone and are the most abundant single rock type within the sedimentary section.

Although graywackes are normally gray or dark in color (the term "graywacke" means a gray grit), nevertheless

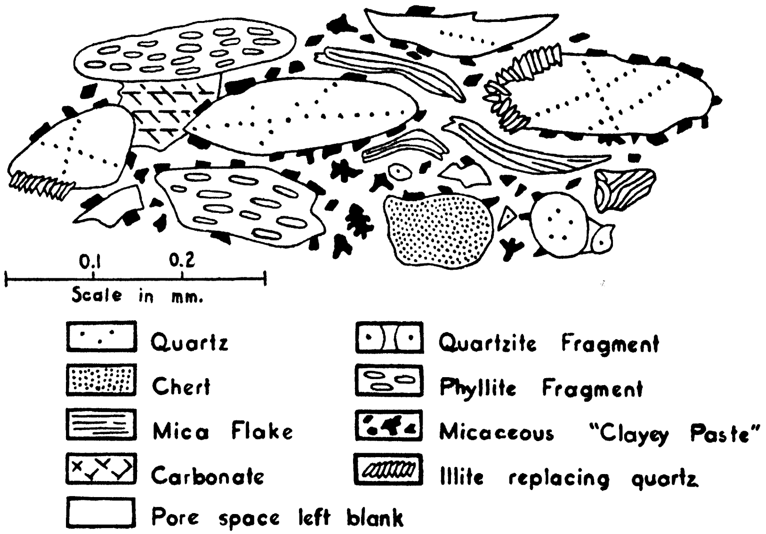



FIG. 13.—Schematic representation of a typical low-rank graywacke (magnification 125×). The diagram is an "exploded" one, i.e., the constituents have been slightly pulled apart for the sake of clarity.

Structurally, the graywackes occur as very large sedimentary bodies, thick and extensive, although the individual sandy or shaly members may be very thin and rapidly changing. Extreme lensing, channeling, and wedging-out are notable. Graywackes are formed during the stage of moderate deformation of the diastrophic cycle, the stage of subsidence, filling, and very early deformation of the geosynclines. Low-rank or common, nonfeldspathic graywackes are related to broad and gently subsiding geosynclines (miogeosynclines). High-rank or feld-

they can also be formed under the oxidizing conditions of a continental environment and in this case may be red.

ARKOSE SERIES

The arkosic series of sediments can be popularly defined as "ashy, light-gray (or red) dirty sands with much feldspathic material."

An arkose (fig. 14) usually contains one or two major detrital clastic end-members: grains derived from the rapid erosion of a granite (granitic detritus), in some cases mixed with a finer-grained

matrix of clay minerals (kaolinite-montmorillonite type) and frequently much iron oxide. Fragments of metamorphic rocks occur as subordinate impurities.

Chemical end-members are not very common but, where present, may range from carbonates (usually found) to gypsum and salt (rare).

The grains of an arkose are coarse to medium (coarser than graywackes and quartzites) and are generally angular but may often be extreme; they assume a prismatic shape.

Arkoses are related to the stage of maximum deformation of the diastrophic cycle (orogenic stage) and are formed either during or immediately after that stage. They normally follow the closing of a geosyncline and may be contemporaneous with large-scale block faulting. Many, if not most, arkoses are of continental origin and thus may be red in

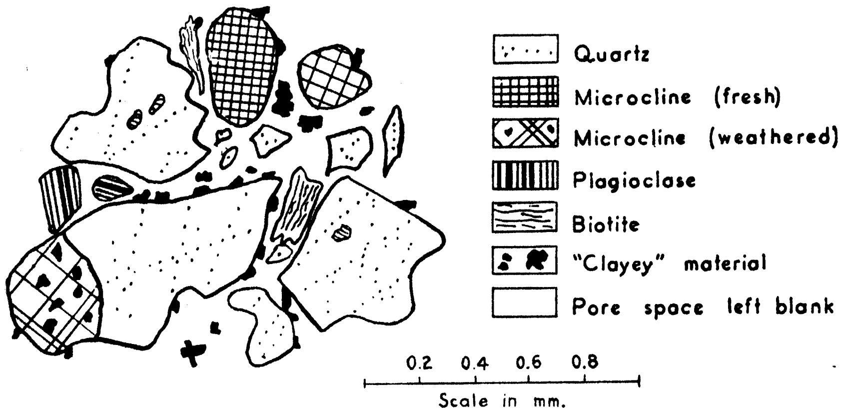



FIG. 14.—Exploded schematic representation of an arkose (magnification 40×)

somewhat more subequant in shape than they are in graywackes. The average grain size of a typical arkose is 1.0–0.18 mm.

The fabric pattern of the average arkose is a mixture of granitic detritus held together by clay, or less frequently by calcite, commonly loaded with iron oxide.

Structurally, arkoses occur as very thick sedimentary bodies, with extremely thick individual members. (The Triassic arkose is 16,000 feet thick in Connecticut.) Lensing and channeling are common, but less so than in the graywackes. Many arkosic bodies are geographically limited in extent, and thus, since their thickness

color. The Triassic arkose of the eastern United States can be considered the typical representative of the arkose series.

### SHALES AND SILTSTONES

The fine-grained members—shales and siltstones—of each of the three major (and four minor) petrographic series discussed above are in reality mechanical mixtures made up of approximately 50 per cent of a fine silt which has essentially the same composition as the corresponding sandstone member (discussed above), roughly 35 per cent of a clay or fine mica fraction of a type which differs for each of the four petrographic subtypes, and 15 per cent of a series of

chemical cements and authigenic minerals typical of shales in general (such as phosphates and titanium oxides). These are very generalized and somewhat preliminary figures. The study of shales is difficult at the megascopic level and usually requires the use of microscopic, X-ray, and thermal analysis methods. The four types of shales are the quartzose, micaceous, chloritic, and kaolinitic-feldspathic classes. However, it is not always possible to make such a distinction with the hand lens, although in some instances it can be done.

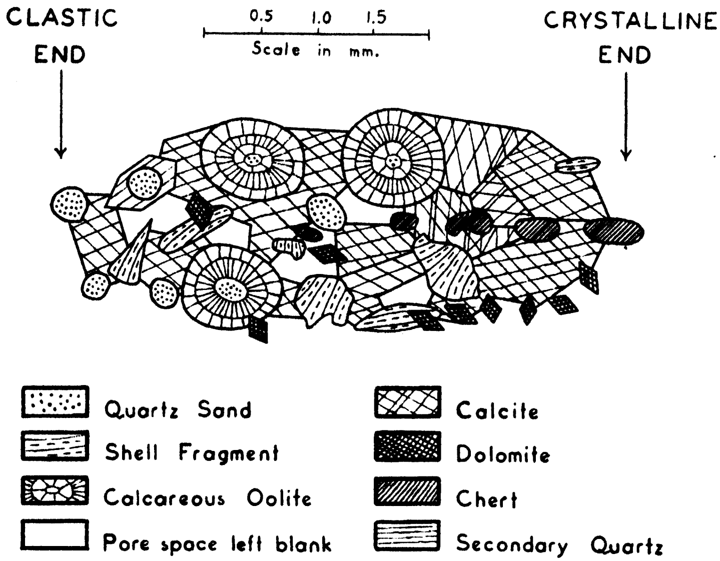



FIG. 15.—Schematic representation of an idealized chemical rock (magnification 16×) showing the passage from a sandy, clastic, oölitic limestone at the left into a densely crystalline cherty and dolomitic limestone at the right.

### CHEMICAL ROCKS (QUARTZITE-LIMESTONE SERIES)

The chemical rocks (limestone, dolomites, cherts, gypsum, salt, phosphates, etc.) usually carry a certain load of detrital impurities which frequently make it possible to fit them into one of the three major petrographic types discussed above.

Most of the chemical rocks (carbonates) are related to the quartzitic series of sediments, and most chemical rocks are formed during the quiescence stage of diastrophism, which is the stage during which epicontinental deposition is at its peak—hence the proposed term "quartzite-limestone series," which indicates a normal and very common association between these two rock types (fig. 15).

### DISTRIBUTION OF SEDIMENTARY TYPES IN THE STRATIGRAPHIC COLUMN

On the basis of a rather comprehensive petrographic and stratigraphic survey of the sediments of the United States, reinforced by the study of several fairly representative rock suites from western Europe, India, Alaska, Burma, and South America and a review of the

meager literature available on the subject, a preliminary estimate of the relative volumetric distribution of the various sedimentary series has been made.

These data are presented in table 4. The figures are believed to be of the correct order of magnitude, but no claims as to their final precision can be made at the present time.

genetically active medium. As stated before, none of these methods is capable of yielding reproducible objective results or of conveying exactly in unambiguous petrographic terms what is meant by the name employed. Furthermore, the description and definition of texture *must* also be based on the use of objective and quantitative terms.

TABLE 4

VOLUMETRIC DISTRIBUTION OF SEDIMENTARY ROCKS IN THE EARTH'S CRUST

| | Quartzite Series | Graywacke Series | | Arkose Series |
|---|---|---|---|---|
| Conglomerates and sandstones | Orthoquartzites 9 per cent | Low-rank graywackes 14 per cent | High-rank graywackes 4 per cent | Arkoses 13 per cent |
| Siltstones and shales | Quartzose shales 2 per cent | Micaceous shales 21 per cent | Chloritic shales 6 per cent | Kaolinitic shales 13 per cent |
| Chem. rocks | Limestones, dolomites, cherts, etc. 18 per cent | | | |

## PART 3. PRINCIPLES OF APPLIED PETROGRAPHIC NOMENCLATURE AND USE OF THE PROPOSED CLASSIFICATION TABLE

In practical matters the end is not mere speculative knowledge of what is to be done, but rather the doing of it.—ARISTOTLE.

### PRINCIPLES OF IDENTIFICATION

The adequate description of any rock requires the naming of its mineral constituents and the definition of its texture. The degree of precision attained depends upon the character of the tools employed: megascopic or microscopic examination.

Proper identification of any rock is *impossible* without the *correct* identification of the principal mineral constituents. In such an incomplete case the classification at best can be based upon texture, bulk chemical or mineral composition, suspected mode of origin, or

### THE PETROGRAPHIC CLASSIFICATION OF IGNEOUS ROCKS: THE MAIN NAME AND ITS QUALIFIERS

A simple and petrographically adequate terminology is used for the classification of igneous rocks. This system is based on two major types of texture (crystalline and glassy as against clastic) and four minor types (crystalline-phaneritic, crystalline-aphanitic, glassy, and, finally, fragmental or clastic) as plotted against two major groups of diagnostic essential minerals (feldspathic versus ferromagnesian) and eight minor groups (quartz plus orthoclase, etc.).

A certain combination of texture and composition determines what is known as a "major" rock type. The name of such a rock type implies both a certain definite composition and a certain typical major texture. For instance, a granite is

a phanerite made of quartz and orthoclase.

These basic rock names are further qualified at the megascopic level by certain subtextures and by the presence of varietal minerals. The subtextures of igneous rocks are based on the degree of homogeneity (porphyritic or nonporphyritic) and on the grade size (coarse, medium, or fine) within the phaneritic textural class. Furthermore, color is given if necessary.

Hence an adequate description of an igneous rock at the megascopic level reads something like this: (1) a white, medium-grained, hornblende granite, or (2) a pink, subporphyritic, two-mica granodiorite, or (3) a greenish, very coarse, augite diorite-porphyry, or (4) a reddish quartz-felsite.

The final term or main name (granite, diorite, felsite, etc.) indicates both the dominant mineral composition and the prevailing texture. The other terms qualify the name and exactly describe the specimen. Certain main names, such as "pegmatite," may have a definite genetic connotation. A similar system can be applied to sediments.

### THE PETROGRAPHIC CLASSIFICATION OF SEDIMENTS: THE MAIN NAME AND ITS QUALIFIERS

Just as in igneous rocks, it is possible to apply the following typical basic standardized descriptive sequence to sedimentary rocks: color, subtexture varietal minerals and cement, and, finally, main name.

However, sediments, as contrasted with igneous rocks, are usually characterized by a strong development of structures and in some cases may also be characteristically connected with a certain transporting medium. Hence, *if necessary*, these two additional properties may be introduced into the descriptive sequence in the following way: color—structure—subtexture—varietal minerals and cement-genetic affinity—main name. Some main names may have a definite genetic significance, such as tillite.

Examples of the classification are as follows:

1. Dark-gray, pebbly, micaceous graywacke.
2. Light-gray, thinly laminated, poorly consolidated, calcareous graywacke.
3. Red, cross-bedded, silty, siderite-bearing graywacke.
4. Red massive conglomeratic arkose.
5. White, fine-grained, rounded, dolomitic quartzitic sandstone.
6. Orange micaceous marl.
7. Pink massive feldspathic silt (or loess).
8. Gray, cross-bedded, coarse-grained, feldspar-bearing sandy dolomite.

### THE MAIN NAME

The main name of a sediment should reflect both the dominant texture and the typical mineral composition. Unfortunately, since the descriptive petrographic terminology of sediments is still in its infancy, there are not enough specific terms (not even in the medium-grained detrital class) to indicate even the principal possible combinations of texture and composition.

It has been thought best to refrain, for the time being, from coining new and fancy rock names, although a tenable terminology can easily be evolved from the present classification table. This agreeable (from an author's point of view) occupation has been left to future students of the problem.

Hence some of the commoner names are used (some slightly redefined), in combination with qualifying terms based on additional information showing in greater detail the rock's composition (like the term "quartz-diorite" among

igneous rock names). This is somewhat awkward but precise.

The main name, therefore, conveys both composition and texture; for example, an arkose is a medium-grained detrital rock made of quartz and feldspar. Similarly, a sandy arkosic limestone is a chemical rock of clastic texture containing more than 50 per cent of calcite (much of it in rolled grains), with a sizable amount of actual quartz and feldspar grains; or a quartzose marl is a fine-grained detrital rock made of very fine-grained, megascopically unrecognizable, aluminum silicates, with a large amount of calcareous cement and carrying recognizable quartz grains. Finally, a quartz conglomerate is a coarse-grained detrital rock consisting of rounded quartz pebbles, whereas an arkosic breccia is a coarse-grained rock, consisting of angular quartz and feldspar pebbles. If this breccia is demonstrably of alluvial-fan origin, the term "arkosic fanglomerate" is applicable; if it is certainly glacial in origin, the term, "arkosic tillite" may be used.

The main name is modified by color, subtexture, character of bonding material, presence of important varietal minerals, and, if necessary, structure and genetic agent.

### QUALIFICATION BY COLOR

The best and safest procedure in describing the color of a sediment is to use a recognized color chart. Since in practice this is generally very difficult, it is best to limit one's self to fairly objective and commonly understood names, such as those found in the basic colors of the spectrum (violet, blue, green, yellow, red), reinforced by the following additional terms: brown, olive, orange, pink, purple, white, gray, and black. Furthermore, it is permissible to combine any two colors (such as bluish-green or greenish-blue) with the second term being the dominant one. Finally, it is also possible to preface each term by the words "light" or "dark" to indicate the intensity of the neutral background (frequently kaolin in the one case, organic matter in the other). Subjective terms, such as "chocolate brown," should be avoided.

### QUALIFICIATION BY SUBTEXTURE: GRADE SIZE AND ANGULARITY

#### IN DETRITAL ROCKS

The following scheme is proposed for detrital sediments:

1. In the coarse-grained rocks the main name is the *XYZ* conglomerate, breccia, tillite, or fanglomerate. The *XYZ* refers to composition. This main name should be prefaced by the term "boulder," "fine," "sandy," "silty," or "clayey" if necessary. When normal and uncomplicated (i.e., 4–64 mm. in diam. with no finer admixtures), the rock is just a normal unqualified conglomerate, breccia, tillite, etc., of a definite composition.

2. In the medium-grained classes the main names, such as quartzite (or quartzitic or quartzose sandstone), graywacke, or arkose should be prefaced by the qualifiers "conglomeratic," "pebbly," "silty," or "clayey" if necessary. Additional terms implying a definite grain size, from very coarse to very fine (if correctly determined) should be added here.

3. In the finer-grained detrital rocks the main names "siltstone" (properly determined as to mineral composition) or "loess" are to be prefaced when necessary by the qualifier "sandy," whereas the main names—shale, clay, or marl—may be prefaced again, when necessary, by the term "silty."

The angularity or, conversely, the rounding of the constituents is strikingly seen in the coarser clastics and is such an obvious property that it should be incorporated in the name. A breccia or fanglomerate or tillite (as contrasted with a conglomerate) should contain no less than 25 per cent of definitely angular fragments (i.e., fragments with sharp edges).

In the medium-grained clastics angularity is a less noticeable property. As a rule, the constituents of all graywackes and almost all arkoses are relatively angular, whereas those of most quartzites are rounded. Pseudo-angularity produced by secondary silica overgrowths on the grains of quartzitic rocks should not be mistaken for original angularity.

The term "grit" is considered superfluous. It usually refers to a pebbly graywacke or a pebbly arkose, rocks in which the grains are normally angular, but where the increased grain size makes this angularity more obvious to the casual observer.

Where observable, the rounding or angularity should be noted by using approximate qualifiers before the main name.

#### IN CHEMICAL ROCKS

Two main textures are possible in the chemical rocks: clastic or crystalline.

1. In the sandy chemical rocks (i.e., those containing over 50 per cent of clastic silicates) the main name, such as sandy limestone, sandy dolomite, oölitic chert, etc., should be prefaced by the terms "coarse," "medium," or "fine" to indicate grain size.

2. In the pure chemical rocks (i.e., those with less than 5 per cent of clastic silicates) the texture is also probably clastic if over 10 per cent of the rock consists of obvious fragmental material of any kind, such as fossil shells. In such a case the same terminology as above may be applied. The main name in this case will be clastic (or arenitic) limestone and so on, and these main names will be modified by the following subtextures: coarse, medium, or fine.

3. If the texture is crystalline, the main name is limestone, dolomite, gypsum, etc., and the qualifier is "coarsely grained," "medium-grained," or "fine-grained." If the crystallinity is to be emphasized or is conspicuous (as is usually the case with coarsely crystalline rocks, which, again, are easier to describe than finer-grained ones), then the term "coarsely crystalline" may be used.

#### QUALIFICATIONS BY SUBTEXTURE: BONDING AGENTS

1. If the detrital rocks are loose (unconsolidated) and lack bonding, the terms "gravel," "boulder bed," "sand," "silt," or "clay" should be used.

2. If the detrital rocks are consolidated by a nonchemical matrix (either produced by simple adhesion of the original finer-grained detrital constituents or by their later reorganization), then the main name is used without additional qualification if the matrix is of the normal mineral composition for that particular rock. For instance, the normal matrix for a graywacke is micaceous or chloritic, whereas the normal arkose possesses an argillaceous, i.e., a kaolinitic matrix, and the name "arkose" presupposes such a matrix. If, on the other hand, the matrix of an arkose is abnormally, but obviously, high in fine micas or in chlorite, then the qualifier "micaceous" or "chloritic" is added to arkose.

A quartzite or a quartzitic sandstone normally has no matrix whatsoever. It is bonded exclusively by chemical cements.

Hence any matrix in a quartzitic sediment is abnormal, and in such a case the

main name, "quartzite," is qualified by the term "chloritic" or "argillaceous." The chances are that such rocks will prove to be very close to graywackes and may turn out to be the so-called "winnowed" graywackes with an abnormally high quartz content (a typical and excellent type of oil reservoir). These borderline and transitional types are going to be discussed in some detail in a forthcoming publication dealing with the genetic aspects of the problem.

If the argillaceous, micaceous, or chloritic material is red in color, the term "ferruginous" *should not be used*, since the bonding effect is produced not by the iron oxide but by the clayey material. Since in this case the ferric oxide acts only as a pigment, the term "red" should be used instead.

3. If the detrital rock is consolidated by a chemical cement, then the main name should be prefaced by the terms "siliceous," "calcareous," "dolomitic," "ferruginous," "glauconitic," or "phosphatic." In less frequent cases the main name may be followed by such expressions as "cemented by gypsum" or "by halite."

The term "ferruginous" is definitely restricted to a chemical cement consisting of ferric oxide. Even then the term is not absolutely satisfactory, and it would be much better to employ the term "hematitic" or "limonitic" instead; but this may be asking too much of the average field geologist without a streak plate. It should be pointed out, nevertheless, that the term "ferruginous" is a gross misnomer when used promiscuously to designate red rocks in general. Indeed, many dark rocks, containing iron in the ferrous state, may have more iron than do some red rocks in which the iron is in the ferric state. In igneous rocks the term "ferruginous granite" is immediately recognizable as a patent absurdity, whereas the term "biotite or hornblende granite" indicates that the iron is there and also how it occurs. The same elementary care in definition should be employed in the description of sediments.

### FREQUENCY OF DISTRIBUTION OF SEDIMENTS IN TERMS OF BONDING MATTER VERSUS DETRITAL MATTER

#### FREQUENCY OF BONDING AND CEMENTING MATERIAL IN DETRITAL ROCKS

In theory any detrital rock can occur with any kind of cement; in practice this is not so. It has been found empirically that certain mineral assemblages in the detrital fractions have a definite tendency to occur with certain types of chemical cements or may lack chemical cements altogether. The genetic reasons for these variations are given in some of the references already mentioned (Krynine, 1942, pp. 537–561; 1943; 1945, pp. 12–22).

The sympathetic or antagonistic associations of certain detrital fractions with certain types of chemical cement are shown by the plus (+) signs when common and the minus (−) signs when rare. These signs indicate the degree of probability for the existence of a given rock with a given type of bonding matter and chemical cement. For instance, the minus (−) sign opposite the $CO_3$ symbol in the feldspathic graywacke column shows that this rock is very rare. Indeed, calcareous high-rank graywackes are almost nonexistent. On the other hand, the plus-minus (±) signs opposite the $CO_3$ symbol in the common graywacke column shows that the probability of its occurrence is better than fair; and, indeed, calcareous low-rank graywackes are frequently found (for instance, the so-called "molasses"—in the petrographic sense—of the European geologists are just that).

The symbols have the following approximate values or order of magnitude:

$\ddagger$ = Always present, about 90 chances out of 100 for its occurrence.
$+$ = Normal type of occurrence, excellent probability of being present, from 70 to 90 chances out of 100.
$\pm$ = Fair to good chances of being present, about 50–70 out of 100.
$\mp$ = Rather poor chances of occurrence, about 20 to 50 out of 100.
$-$ = Very poor chance of being present, less than 20 out of 100.

These percentages refer to frequency of occurrence within the same rock type. For instance, in the case of common low-rank graywackes the plus (+) sign in the *A*, *M–C* tier indicates that of 100 low-rank graywacke specimens perhaps 70–90 per cent will carry as the predominant bonding medium (either alone or in combination with subordinate cements) an argillaceous or rather micaceous chloritic matrix. Furthermore, the plus-minus (±) sign in the carbonate tier of the same graywacke class indicates that of the same 100 low-rank graywacke specimens 50–70 may contain sizable amounts of carbonate cement, and so on for the other bonding media.

The use of plus and minus signs is for the guidance of the student and shows him what the odds are for the possibility of the occurrence of a given rock type and thus serve as a rough check on his identification of the rock.

The use of the plus and minus signs in the finer-grained detrital rocks is similar to that in the coarse and medium-grained class. For instance, the minus sign opposite the $SiO_2$ symbol in the siliceous or opaline shale tier in the low-rank graywacke column indicates that such siliceous shales are rather rare where they carry a coarsely detrital fraction of micaceous composition with no feldspar but are common (+) if feldspar appears in addition to mica and chlorite (feldspathic graywackes column).

These values are admittedly very general approximations only, but their order of magnitude is, on the whole, believed to be correct.

#### FREQUENCY OF SUBORDINATE DETRITAL MATERIAL IN CHEMICAL ROCKS

Since the chemical fraction is dominant in the chemical rocks, the use of the plus and minus signs in this category is reversed, and they are employed to indicate the probability of occurrence of a certain type of detrital fraction within a given background of chemical precipitates.

For instance, the double plus ($\ddagger$) signs in the quartzite column of the sandy limestone class indicates that sandy limestones generally contain quartz (or chert) grains and, conversely, that they are almost nonexistent with a detrital load corresponding to that of a high-rank graywacke (−) but that they have a better possibility of carrying an arkosic load (∓) and a very good possibility of carrying a low-rank graywacke type of detritus (+).

The plus and minus signs left and right of the oölitic subcaptions show that the probability of oölitic occurrence increases as the silicate detritus assumes a more purely quartzose character.

#### QUALIFICATIONS BY VARIETAL AND ACCESSORY MINERALS

The essential minerals which determine the main name of a sediment are specifically listed under the heading of "Composition." Within the detrital rocks these essential minerals are quartz-chert for the quartzitic series; quartz, chert, mica, micaceous clays, and rock fragments for the graywacke series in general and also feldspar for the high-

rank graywacke subseries; and quartz, feldspar, and nonmicaceous normal clay minerals (i.e., kaolinite and bauxite) for the arkosic series of detrital rocks.

Any other mineral not specifically mentioned in this list becomes a varietal mineral if it exceeds 1 per cent in amout and hence becomes a major constituent that is readily recognizable with a hand lens. The terminology employed to indicate the presence of such varietal minerals should, theoretically at least, be different from that used to designate the occurrence of chemical cements.

A system employing a terminology such as "gypsum-bearing" or "mica-bearing" or "garnet-bearing" would have the advantage of being entirely clear but would be awkward and cacophonous and hence undesirable. As a compromise between scientific and linguistic clarity the following system is tentatively proposed:

*a*) Adjectives ending in *-ic* or *-ous* should be used for varietal minerals when no confusion is possible with chemical cements. For instance, in "micaceous arkose" it is clear that mica is a varietal detrital mineral (possibly in the matrix) rather than a chemical cement.

*b*) If confusion is possible, then the less elegant but more precise term "mineral-bearing" is to be employed. An example would be "glauconite-bearing dolomitic graywacke."

Under "accessory minerals" are included mostly the so-called "heavy" minerals. Generally these minerals cannot be studied at the megascopic level. Any mineral which occurs in amounts of less than 1 per cent in a sediment is considered to be an accessory and is not mentioned in the megascopic definition of the rock. Its presence, if determinable megascopically, is mentioned in an additional sentence when proceeding with the detailed description of the rock.

QUALIFICATIONS BY STRUCTURE

Whereas in igneous rocks a massive structure is the rule and layered or banded structures are the exception, the opposite is true in sediments. Two possibilities may exist: (1) the structure of a sediment is simple and can be defined in one or two words (cross-bedded, thinly laminated, varved, massive); or (2) the structure is complex and needs several words or an entire sentence for its definition.

In the first case the definition can be incorporated in the description preceding the main name, and its place will be between color and subtexture (for instance, a red, cross-bedded, pebbly arkose).

In the second case the definition should follow the main name and should be prefaced by the word "showing." For instance, a "dark-greenish-blue, silty, high-rank graywacke, showing well-developed cyclic graded bedding in 4-cm. bands with incipient and semimicroscopic cross-bedding within the coarser portion of each 4-cm. layer." If the description of the structure begins to be loo long, it can be separated into a sentence of its own following the basic definition of the rock.

The presence of fossils and recognizable organic remains is considered to be a structure and should be recorded at this state (example: white, coarse, cross-bedded, fossiliferous, dolomite-bearing quartzite).

GENETIC AND ENVIROMENTAL TERMS

Many rock types are produced by certain specialized sets of processes (frequently and somewhat loosely referred to as "environments"). This combination of processes may produce a certain typical combination of composition and texture and possibly structure in the resulting rock. The nomenclature of these specialized sets of processes ranges from

such generalized terms as "marine" or "continental" to much more definite terms, such as "eolian," "glacial," or "of alluvial-fan origin," etc. In the last three instances the products of these processes have been frequently called "loess" (a somewhat loose usage), "till," and "fanglomerate." Theoretically and in the abstract, any kind of mineral composition should be possible for a till or a fanglomerate. Actually this is not so because genetic processes, formative conditions, and so-called "environments" do not exist in a geologic vacuum but operate against a certain dominant diastrophic background which not only determines the probable relative intensity and effectiveness of any one process but also the type of material that it will have to work with.[1]

For instance, the rigorous climatic or topographic conditions which produce a fanglomerate, a tillite, or a loess (deflated material of a glacier or of a dried-up fluvial basin) are all very typical of the large-scale emergence of continents and of periods of relatively intense orogeny and hence are characterized by arkosic sediments, possibly diluted by some graywacke-like material. The validity of this conclusion has been pragmatically verified by a petrographic study of numerous assorted loesses, tillites, and fanglomerates. For these reasons loess and fanglomerates are tentatively defined not only as eolian or alluvial-fan type sediments but also as highly feldspathic ones.

Tillites are also normally defined as basically arkosic and hence in special cases should be qualified as graywacke-tillites. Hence these genetic terms are elevated to the rank of main names, characterized by composition, texture, and also in this case formative agents. An igneous analogy is the term "pegmatite," which also has a definite genetic meaning.

Other genetic terms, such as "eolian," "flood-plain" (deposit), "glacial," "continental," "marine-beach" (deposit), "desert" or "desertic" (deposit), "paludal," "lacustrine," etc., if considered necessary and if conclusively known to be *correct* for the particular sediments under study, can be used as qualifiers to redefine the main name. These terms should be placed before the main name if they are brief or after the main name if they are lengthy. In the latter case the main name should be followed by a statement that the specimen is a clayey graywacke-breccia (or boulder clay) probably of late glacial origin.

Such dubious terms as "silicified" or "recrystallized" (both great favorites in stratigraphy) and similar pseudo-genetic terms which imply a knowledge that the observer at the megascopic level does not (and cannot) possess, should be avoided. Use, instead, the terms "siliceous" or "crystalline," which are correct, truthful, and adequate.

## USE OF PROPOSED TERMINOLOGY AND CLASSIFICATION TABLE

In summary, the following step-by-step procedure is suggested for the adequate megascopic description and classification of a sedimentary rock:

1. The rock is classified as a detrital or chemical rock, depending on the relative proportion of detrital material (in practice almost entirely restricted to *clastic* silicates) and chemical material.

2. The rock is classified further as belonging to the quartzite, graywacke, or arkosic series, depending upon the com-

[1] The principles behind this concept are explained in a general way in the references cited earlier (Krynine, 1942, pp. 537–561; 1943; 1945, pp. 12–22), and are to be described in specific detail in a forthcoming publication.

position of the clastic silicate fraction. This division when dealing with medium-grained clastic rocks (detrital rocks or sandy chemical rocks) can practically *always be accomplished successfully by a moderately well-trained observer* and has been successfully carried out without undue difficulty by upperclassmen and even Sophomores working under the writer's guidance at the Pennsylvania State College.

3. Such a division may not be possible when dealing with fine-grained clastic rocks (detrital rocks or sandy chemical rocks) or with pure chemical rocks with no recognizable silicate fraction. Then the terms "limestone," "siltstone," and "shale" are to be used in the same way as the term "felsite" is used in igneous rocks.

4. If the rock is a detrital one, after placing it in the proper series, determine whether it is bonded by a fine-grained matrix or a chemical cement. This can be accomplished, again without undue difficulty, by observing the clastic and grained character of a matrix as compared with the densely crystalline character of a chemical cement. At this stage the plus and minus signs will serve as a guide to the probability of the identification's being a correct one.

5. If the rock is chemical, note whether it is sandy (over 5 per cent of clastic silicates) or pure. If it is pure, determine whether the texture is arenitic (clastic) or crystalline.

6. If the chemical rock is sandy, determine the detrital series to which the sandy fraction belongs. This can be done with little effort in many, or most, sandy chemical rocks if they are medium or coarser grained. If the chemical rock is pure, this may prove to be impossible; but the attempt should always be made. Again the plus and minus signs will serve as a guide. If necessary, dissolve a small amount of the rock in acid and examine the insoluble residue with a hand lens.

7. Proceed with the complete identification and description of the rock as suggested in the preceding paragraphs by following the scheme of: color—structure—subtexture—varietal minerals—cement and bonding (or, conversely, detrital material in a chemical rock)—genetic affinity—main name.

## CONCLUSIONS

### SCIENTIFIC SIGNIFICANCE

In the present age of specialization there has been a tendency among field geologists not to bother unduly with rock specimens and to leave their detailed description to petrographers. As a result, a comparison of many recently published stratigraphic and areal geologic studies shows that, in so far as descriptions of rocks (which, after all, are the primary building blocks of all geological work) are concerned, these recent publications are less detailed—and frequently less correct—than similar descriptions published over a hundred years ago. For instance, the rock descriptions found in Percival's classic monograph on the geology of Connecticut, published in 1842, will compare favorably, both in precision and in accuracy, with most modern nonspecialized megascopic treatments of sediments.

Since all rocks are observed in the field and most of them never get past this stage of hand-lens description, it follows that a considerable improvement in megascopic standards would be most useful. In addition to the satisfaction of a duty well done, such improvement would provide, among other things, adequate material (so-called "sedimentary evidence") for a much better interpretation of paleogeographical and structural

events, an interpretation which to a large extent should be carried out in the field.

### PRACTICAL APPLICATIONS

At the present time approximately 50–60 per cent of graduates in geology go into the petroleum industry. It is not an exaggeration to say that during the first few years of their careers, at least half of their time will be spent "sitting on wells," i.e., examining drill cuttings and drawing the necessary geological conclusions, which—if wrong—may be extremely expensive. On that basis it follows that the odds are strong that at least one-third of the total work done by the average geologist for five years or so following graduation will be spent on the equivalent of an applied course in elementary hand-specimen petrology, and specifically on the sedimentary portion of it. Hence, to put it mildly, the standards of such work—again in respect to both precision and accuracy—ought to be greatly improved.

Search for the so-called "stratigraphic traps" is mostly a petrographic undertaking, which, if accomplished in the field or in the average oil-company laboratory, again resolves itself into the interpretation of megascopic data. Stratigraphic traps are essentially a function of the relations existing between grains, on the one hand, and either matrix (the common detrital sand-shale lensing) or cement (equally common changes in chemical constituents), on the other. The characters of these changes are quite different in different rock series, and hence an adequate working knowledge of simple sedimentary elements and the ability to recognize them rapidly in the field are desirable. Finally, studies by the writer have shown that the three major sedimentary series are characterized by three entirely different types of oil fields and that the methods of prospecting—and exploitation—which are successful in one series may fail completely in the other.[2]

This should come as no surprise to the igneous petrographer and economic geologist, who know that tin deposits are related to granites and nickel deposits to gabbros and dunites. Similar genetic relationships exist in the distribution of different types of petroleum deposits. Such genetic relationships can be understood best when the description and classification of sedimentary rocks reaches a degree of precision at least equal to that proposed in the present article.

## REFERENCES CITED

De Ford, R. K. (1944) Rock colors: Am. Assoc. Petroleum Geologists Bull. 28, pp. 128–137.

Krynine, P. D. (1942) Differential sedimentation and its products during one complete geosynclinal cycle: Annals of the First Pan-Am. Cong. of Min. Eng. and Geology, vol. 2, pt. 1 (Geology), pp. 537–561.

——— (1943) Diastrophism and the evolution of sedimentary rocks (syllabus outline of lecture), Distinguished Lecture Comm. of Am. Assoc. Petroleum Geologists.

——— (1945) Sediments and the search for oil: Producers Monthly, vol. 9, no. 3, pp. 12–22.

[2] The seismograph, for instance, works very well in the orthoquartzite-limestone series but frequently does not work at all in the graywackes.

# REVIEWS

*Second Symposium on the Age of the Saline Series in the Salt Range of the Punjab Held at Udaipur on 27 and 28 December, 1945, under the Joint Auspices of the National Academy and the Indian Academy of Sciences.* (*Nat. Acad. Sci. India Proc.*, sec. B, vol. 16, [April, 1947].) Pp. 1+257+ many illus. Rs. 15 to nonmembers abroad.

The so-called "Saline series," or salt marl, occurs at the base of the sedimentary sequence of the Punjab Salt Range and has long been the subject of one of the most interesting controversies in Indian geology. At most places these beds underlie the unfossiliferous Purple sandstone (believed to be Cambrian because it underlies fossiliferous Cambrian strata) or the Talchir boulder beds (late Paleozoic tillite). Stratigraphic position suggests that the Saline series is of early Cambrian or pre-Cambrian age. Nowhere has it been observed in unquestioned, undisturbed contact with older rocks.

Generally similar salt-bearing beds occur in the Kohat region west of the Indus River and are recognized by all to be of late Eocene age. Outcrops of these deposits are separated by only 17 miles from the salt marl of the Salt Range. This fact, the generally "young" appearance of the Saline series, and the presence of oil shale and shows of oil within it have naturally led to the conclusion that the salt-bearing beds of Kohat and the Punjab are of equivalent age and that the latter owes its apparent stratigraphic position to the overthrusting of more ancient strata.

These divergent views regarding the age of the Saline series and the fundamental structure of the Salt Range have been debated for many years. More recent investigations have furnished evidence interpreted as being favorable to both contentions. Many fragments of angiosperm and gymnosperm plants and insects of undoubted post-Paleozoic age have been recovered from the rock salt, dolomite, and oil shale of the Saline series. On the other hand, field studies have failed to establish a widespread plane of thrusting above these strata, and at many places the stratigraphic sequence appears to be normal and undisturbed.

The present symposium is the second to consider this problem. It followed a field excursion to the Salt Range, where a number of critical exposures were carefully examined and sampled.

The principal proponents in the controversy at the present time are Professor B. Sahni, of the University of Lucknow, and Mr. E. R. Gee, of the Geological Survey of India, who favor late and early ages of the salt marl, respectively. Professor Sahni discovered fossil fragments in the rock salt and has been instrumental in interesting several other investigators in their study. Mr. Gee has engaged in extensive field work in the Salt Range and has mapped several parts of it in detail. He formerly indorsed the view that the Saline series is Eocene, but field evidence forced him to change his mind. The positions taken by these gentlemen in general reflect the opinions of paleontologists, on the one hand, and stratigraphic and structural geologists, on the other, although among the latter there are a few important exceptions. Each side maintains its position confidently and tries to explain away conflicting evidence presented by the other. On the whole, however, the case presented by the paleontologists is the more convincing because their findings do not appear to be as liable to errors of observation and interpretation.

The evidence of the fossils has been countered in two ways. First, it is suggested that they are not naturally indigenous to the salt-marl strata but were introduced much later into plastic or porous beds. Although they are found in every sample studied, this explanation would appear to be possible if they were confined to the rock salt. However, they also occur in dolomite and in fine-grained oil shale, some of which consists of more than 50 per cent organic material. All geologists who have studied the area admit that the oil shale is an undoubted original constituent of the Saline series.

All who have examined the organic material agree that it includes fragments of angiosperm and gymnosperm plants and insects that must be of Cretaceous or younger age. If they are original constituents of the rocks—and it is almost inconceivable that they could have been

introduced into the oil shale—the age of the strata is established as post-Paleozoic. Gee's unconvincing secondary defense is the suggestion that these fragments represent an early flora of much more advanced type than anything previously found in beds of comparable age.

The proponents of an Eocene age for the salt marl counter the claim of the field geologist that no widespread thrust occurs above the Saline series by reminding them that thrusting is evident at some places and that the presence of thrust planes unreflected by any recognized local deformation has been established by geologists in several other parts of the world.

The printed symposium consists of nineteen papers, several of them quite inconsequential, and a section on discussion. Seven of the papers were read before the conference, eleven are by authors unable to attend but were briefly reviewed at that time, and three are subsequent contributions. Four papers are devoted principally to descriptions of fossil fragments obtained from the Saline series. Two papers consider theoretical aspects of Salt Range tectonics and support opposite conclusions. Among the more interesting and important geologic contributions are the following: "Microfossils and the Salt Range Thrust," by B. Sahni; "Some Tectonic Aspects of the Problem," by E. Lehner; "A Note on the Saline Series of North-western India," by M. S. Krishnan and N. K. N. Aiyengar; "Further Note on the Age of the Saline Series of the Punjab and of Kohat," by E. R. Gee; and "Further Notes on the Age of the Salt Range Saline Series," by E. S. Pinfold.

J. M. W.

---

*An Introduction to Crystallography.* By F. C. Phillips. New York 3: Longmans, Green & Co., 1946 (July 2, 1947). Pp. ix+302; figs. 500. $6.50.

Part I, constituting three-fourths of this handy volume, is devoted to morphology; of the remainder, two chapters (x and xi) deal with space lattices and space groups and their symmetry, and a final chapter covers crystal habit. Physical (including optical), chemical, genetic, and X-ray crystallography are not treated. The nine chapters of Part I cover the study of crystals by the methods of single-circle reflection goniometry and the stereographic projection.

Chapter iv is a general study of the seven crystal systems in the order cubic, tetragonal, orthorhombic, monoclinic, triclinic, hexagonal, and trigonal; it deals only with the holosymmetric classes (also the pyrite class). The first third of the book could thus serve for an elementary course based on the stereographic projection (chap. ii), with brief treatments of goniometry and crystal drawing (axial cross). The remainder of Part I includes chapters on goniometry, the thirty-two classes (covered in the order of increasing symmetry), composite crystals, mathematical relationships, and crystal drawing. This sort of double coverage of crystal morphology in Part I is handled very successfully. It appears to be pedagogically sound and should result in imparting a good basic concept of the subject.

While it is true that "the main problems of [morphological] crystallography can be solved . . . . by the use of the simple single-circle goniometer," it does not follow that this is the best modern training technique. Thus Barker states that "as much work can be effected by the two-circle instrument in four hours, as in seven hours devoted to single-circle goniometry." Many feel that the two-circle machine, though more complicated to manufacture, is nevertheless much simpler to use. Moreover, the elegance of the associated gnomonic projection (very briefly treated in the present work) is of considerable educational value. This is indirectly suggested on pages 90, 219, and 276, in connection with the reading of face indices, crystal drawing, and the computation of reticular areas. The use of the Goldschmidt method in the new Dana's System is another important factor in favor of this projection.

Chapters x and xi serve as a very good introduction for beginning students to Volume I of the *International Tables for the Determination of Crystal Structures*. The Hermann-Mauguin notation is employed. Space groups are covered in the same order of crystal classes as is given in chapter iv, except that the tetragonal precedes the trigonal. In the final chapter the meaning of the Law of Bravais as generalized by Donnay and Harker is made clear.

The book has remarkably few errors; none of a typographical nature was noted. The definition of a crystal "as a homogeneous solid bounded by naturally-formed plane faces" is applicable only to that minority occurring as euhedrons. The "linear" projection (p. 31) should be called the "euthygraphic"; the former term was first applied to the Quenstedt projection. Hauy's *Traité* has five volumes, not four

(p. 34). The "stroke" in the symbol $2/m$ is called a *virgule* (p. 105). Re-entrant angles (p. 163) are typical of aggregates; a small proportion of these may be twins. The $c < a < b$ rule is not mentioned; thus the orientation of chrysoberyl (p. 170) does not fit Dana's System. The same holds true for topaz (p. 287), which in the new orientation to be adopted in Dana will not be prismatic (as stated on p. 286). Moreover, since it is said not to be holosymmetric (p. 175), the argument on pages 285–86 is vitiated on two counts.

The treatment in this book, greatly aided by numerous line sketches, is logical and lucid; the student should enjoy using it. The problems worked out in detail, together with the careful goniometric instructions, make it a real guide for the study of *crystals* (rather than models). It deserves to succeed, measured (as the author suggests) "by the number of its readers who finally lay it aside and 'throwing off the shackles of the text-book,' set out upon their own crystallographic investigations."

D. J. F.

---

*American Oil Operations Abroad.* By LEONARD M. FANNING. New York and London: McGraw-Hill Book Co., 1947. Pp. 270+vii. $5.00.

Treating his subject from the point of view of an oil company, Mr. Fanning has covered American petroleum operations in foreign countries in a most interesting manner. One must commend the remarkably profuse and well-chosen photographs with which the work is filled.

It has been the reviewer's contention that the public should be apprised of the risks in foreign oil fields. Mr. Fanning has presented effectively the undesirability of attempting such operations without capital running into many millions of dollars. His chapter dealing with political risks stresses the constant danger of expropriation and revision of terms. Economic risks are given almost equal space, with special attention paid to the time lag in dividends. He does not, however, point out the employmental, geological, and technological risks involved.

The chapter on the advantages, to both nations, covers social, educational, and economic factors. Although one may question some of the general premises of his economic discussion, the main points are clearly stated for the benefit of the American public. His picture of educational and social life in foreign oil fields is one unfamiliar to the average citizen. It might well receive more general consideration, with advantage both to individual and industry.

Curiously enough, Mr. Fanning has failed to distinguish between oil companies operating from an American and those operating from a European base, either in regard to their acceptance by, or in their treatment of, the nationals of the concession-granting countries. This reviewer thinks that he has missed an opportunity to point out that, on the whole, the American is better received and has given a better account of himself abroad than has the European.

In the appendixes one finds a vast amount of material; United States and foreign investments, production, price, trade, employment, and other statistics are available galore. They are not, however, broken down by countries, which renders them less usable than they otherwise would be. Furthermore, stress has been laid upon investment and financial considerations rather than upon physical quantities.

For the general reader or as collateral reading in general courses in college the book is admirably adapted. Both Mr. Fanning and the publishers are to be congratulated upon having presented a concise account of the search for oil in foreign lands by American nationals.

From the point of view of the research worker, however, the book has shortcomings. This reviewer is surprised to find a regrettable dearth of information relative to the geology or operating techniques. Nowhere is there a definite statement of the areas held in various countries or the underlying reserves.

H. W. STRALEY III

# Rollin Thomas Chamberlin

*Professor Emeritus of Geology and Editor Emeritus of the Journal of Geology*

BORN OCTOBER 20, 1881
DIED MARCH 6, 1948

AS HAS been aptly said, an institution is commonly but the lengthened shadow of one man. So it was for a quarter of a century with the Journal of Geology. Rollin Thomas Chamberlin gave a considerable part of his life to the Journal and, by his service to that publication, rendered a large service to geologists and the science of geology. His association with the Journal exceeds in length even that of T. C. Chamberlin, its founder and first editor. Rollin Chamberlin was made a member of the editorial staff in 1912; he became managing editor in 1923 and editor in 1929. He remained in this capacity until his retirement on June 30, 1947. Readers and authors alike owe Rollin Chamberlin a great debt for this thirty-five years of painstaking and at times tedious labor. Many readers, perhaps, are unaware of the patient and unending effort contributed gratuitously by the editor in assisting authors in the preparation of their manuscripts for publication.

Rollin Chamberlin's service to the Journal did not consist solely in discharge of his editorial duties. He was also the author of many articles and reviews. His first contribution, "The Glacial Features of the St. Croix Dalles Region," appeared in volume 13 in 1905. His last paper, "The Moon's Lack of Folded Ranges," appeared in volume 53, 1945.

It is the hope of the present editorial staff that the Journal of Geology will continue as a publication worthy of the Chamberlin tradition.

VOLUME 56 NUMBER 3

# THE JOURNAL OF GEOLOGY

*May 1948*

## CAVE-IN LAKES IN THE NABESNA, CHISANA, AND TANANA RIVER VALLEYS, EASTERN ALASKA[1]

ROBERT E. WALLACE
State College of Washington, Pullman, Washington

ABSTRACT

Cave-in lakes resulting from ground caving following the thawing of permafrost have developed in areas underlain by fine-grained sediments in the Nabesna, Chisana, and Tanana River valleys of eastern Alaska. It is suggested that the vegetal cover has an important control over the presence of permafrost and that a cave-in lake is initiated by a break in this cover. Once a lake is formed, the banks retreat at a rate indicated to be of the order of a few inches a year. The recession of lake banks thus enlarges the lake and is responsible for a typical sequence of areal patterns of the cave-in lakes.

### INTRODUCTION

The present study of cave-in lakes in the Nabesna, Chisana, and Tanana River valleys of eastern Alaska was done as part of an investigation of permafrost by the Geological Survey which, in part, was conducted under the sponsorship of the Military Intelligence Division, Office of the Chief of Engineers, U.S. Army. The purpose of the study was to develop criteria to be used in determining permafrost conditions by means of aerial reconnaissance.

Observations were made during October and November, 1945, within a radius of approximately 15 miles of Northway, Alaska, near latitude 63° N. and longitude 142° W., approximately 30 miles west of the Alaskan-Canadian boundary.

### DEFINITIONS

*Permafrost*, a condition in which the ground temperature remains below freezing throughout a considerable number of years. The condition is defined as "dry permafrost" if no ice is present. "Perennially frozen ground," "permanently frozen ground," and "pergelisol" (Bryan, 1946, p. 640) are other terms that are used for permafrost.

*Cave-in lakes*, lakes whose basins result from collapse following volume contraction due to thawing of permafrost. The term "thermokarst" has also been applied to these lakes and alludes to the similarity of origin and appearance of these lakes to karst lakes in areas underlain by limestone. "Kettle lake" and "Kettle-hole lake" have also been used in referring to these lakes.

### GEOLOGIC SETTING

The area studied (figs. 1 and 2) is within a radius of 15 miles of the point at which the Chisana and Nabesna rivers

[1] Published by permission of the Director, United States Geological Survey and Office of the Chief of Engineers, United States Army. Manuscript received November 6, 1947.

join to form the Tanana River. Most of the area is southwest of the Tanana and Chisana rivers. An area immediately underlain by crystalline rocks is northeast of these two rivers.

The broad floodplains of the Nabesna, Chisana, and Tanana rivers lie in the dissected floor of an old lake basin. Remnants of the old lake sediments, including some stream-borne gravels, form terraces, which in some places are surrounded by the floodplains; elsewhere the terraces border the floodplains. The terraces are about 100–200 feet above the present river levels. They have been subjected to wind action, so that large areas are covered by sand dunes, many of which have a relief of 100 feet. The dunes are covered by vegetation and are almost inactive at present.

The Nabesna River flows through a notch in a ridge of bedrock approximately 20 miles from its mouth. A terminal moraine on the south side of the ridge marks the greatest advance of the Nabesna Glacier. The river has incised the moraine to a depth of about 100 feet. Outwash material from the glacier has been carried northward through the notch and has been spread out in the form of an alluvial fan on the dissected floor of the lake basin. The present Nabesna River is slightly incised in the upper part of the fan.

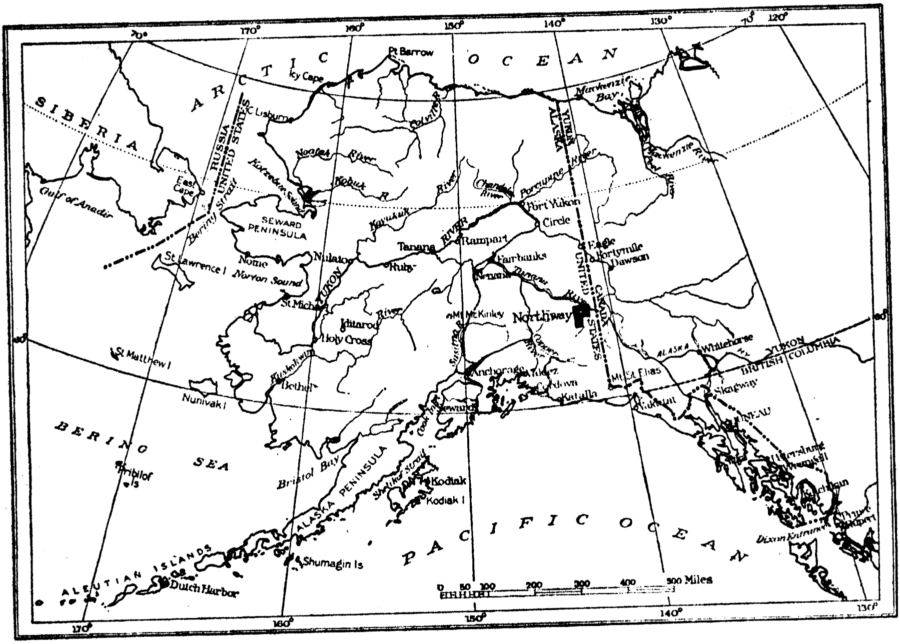



FIG. 1.—Index map of Alaska, showing Northway area

### DISTRIBUTION OF CAVE-IN LAKES

Cave-in lakes have been reported from many parts of Alaska; however, few detailed descriptions are available. Cave-in lakes are manifestations of permafrost; their distribution, therefore, is coextensive with that of permafrost. The southern limit of permafrost is at approximately 60° N. in southwestern Alaska, paral-

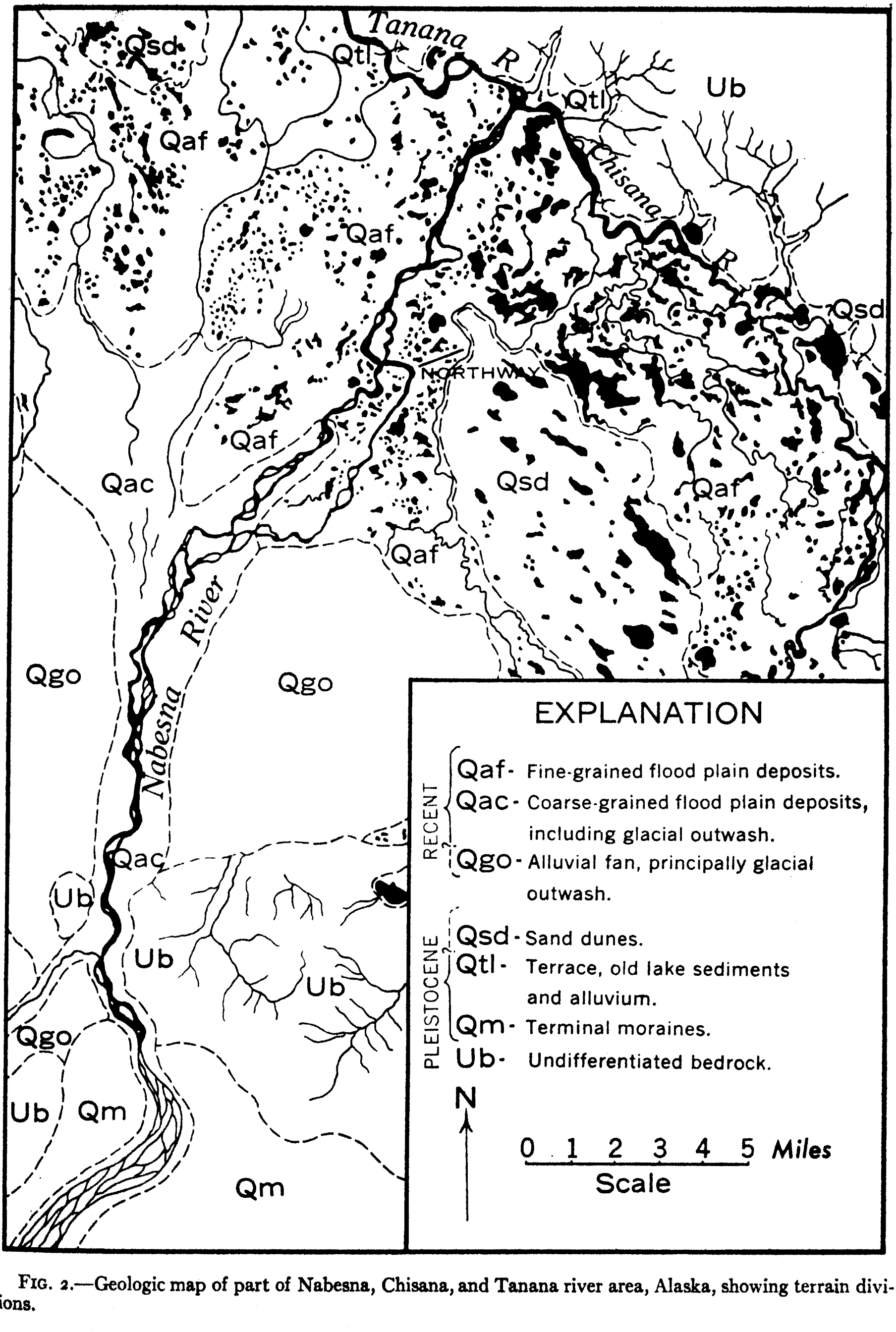



FIG. 2.—Geologic map of part of Nabesna, Chisana, and Tanana river area, Alaska, showing terrain divisions.

lels the coast a few miles inland along the Gulf of Alaska, and is a few tens of miles north of Anchorage, Alaska, at approximately 62° N. (Smith, 1939, pl. 14).

Almost all the cave-in lakes in the area studied are in the low flood plains of the Nabesna, Chisana, and Tanana rivers. Although permafrost was found in most of the other terrain divisions (fig. 2), little evidence of caving was found. For example, although many lakes are present in the sand-dune areas, no caving along the borders of these lakes was noticed. Some caving was found in the terrace areas, but no lakes had developed.

This distribution apparently is controlled to considerable extent by the grain-size of the sediments which are characteristic of the terrain units. Silts and fine-grained sediments change greatly in volume (Taber, 1929, pp. 460–461; Muller, 1945, p. 64) upon freezing and thawing, whereas well-sorted sands and gravels change but little or not at all. The distribution of cave-in lakes, therefore, seems to coincide with the areas in which fine-grained sediments occur within the regions affected by permafrost. Cave-in lakes thereby become an excellent key, recognizable in aerial photographs, for determining permafrost-bearing ground subject to slumping upon thawing.

## CROSS SECTION OF CAVE-IN LAKES

Probing and drilling with hand augers by members of the Geological Survey and extensive drilling and temperature testing in certain areas by engineers of the St. Paul district indicate to a certain extent the distribution of permafrost near and under several cave-in lakes in the floodplain of the Nabesna River.

The cave-in lakes investigated are very shallow and have relatively flat bottoms, most being less than a maximum of 5–10 feet below the surface of the floodplains. The borders of the lakes are typically precipitous, dropping off steeply from the general level of the floodplain to the bottom of the lakes. The vegetal cover left unsupported by caving droops over the precipitous bank and, where the drop is not too great, forms a continuous mat over the cave-in bank and the adjacent bottom of the lake (fig. 3). Where the mat is not flexible or if the drop is too great, large cracks form parallel to the borders of the lake. In extreme cases large land slumps occur.

The depth to the permafrost table under the floodplain ranges from a few inches in places where vegetation affords the best insulation to approximately 10 feet where it provides poor insulation. The depth beneath the average spruce forest cover, typical of the low floodplains, is 1 or 2 feet. The ground in these spruce forests commonly is covered by a mat of sphagnum moss, 2–8 inches thick, low-bush cranberries, *Ledum*, and, in more open areas, dwarf birch, blueberries, and grasses.

The depth to permafrost at the edges of the cave-in lakes increases precipitously (fig. 3). No permafrost was encountered down to a depth of 12 feet, the greatest depth that could be reached with the auger used, in holes that were put down just offshore. No permafrost was found at a depth of 12 feet below the lake surface or 8–10 feet below the lake bottoms in other holes that were put down in the centers of lakes. The character of the permafrost at greater depths below the lakes is not known. The engineers at the St. Paul district have drilled several holes approximately 50 feet deep in the floodplains. These penetrated the entire thickness of the permafrost.

## AREAL PLAN OF LAKES AND SEQUENCE OF DEVELOPMENT

A sequence of cave-in lake development was recognized in the Northway area. This sequence can be divided into four stages, which may be referred to as youthful, early mature, late mature, and old age.

The youthful stage (fig. 4) of the sequence begins when caving of the ground, as the result of thawing, forms a small basin in which water accumulates. As the permafrost in the banks of the depression thaws progressively outward, the banks retreat from the initial point, forming a roughly circular lake which is continually enlarging in area. Some simple circular lakes, characteristic of the youthful stage, are as much as 600 feet in diameter.

The early mature stage (fig. 5) is designated as that in which several simple lakes have joined to form an aggregate. The typical borders of lakes at this stage are "scalloped" because of the original circular shape of the individual lakes.

The late mature stage of the sequence is defined as that in which aggregates of lakes have coalesced with other aggregates and the whole system has been integrated, commonly by intersection of a drainage channel, so that there is drainage between and through the lakes (fig. 6). The outline of the lake at this stage is very irregular because of the complex combination of lakes. The simple circular outline of the original lakes is greatly modified in places. The lakes are flooded and drained repeatedly as the level of the main river fluctuates. Natural levees

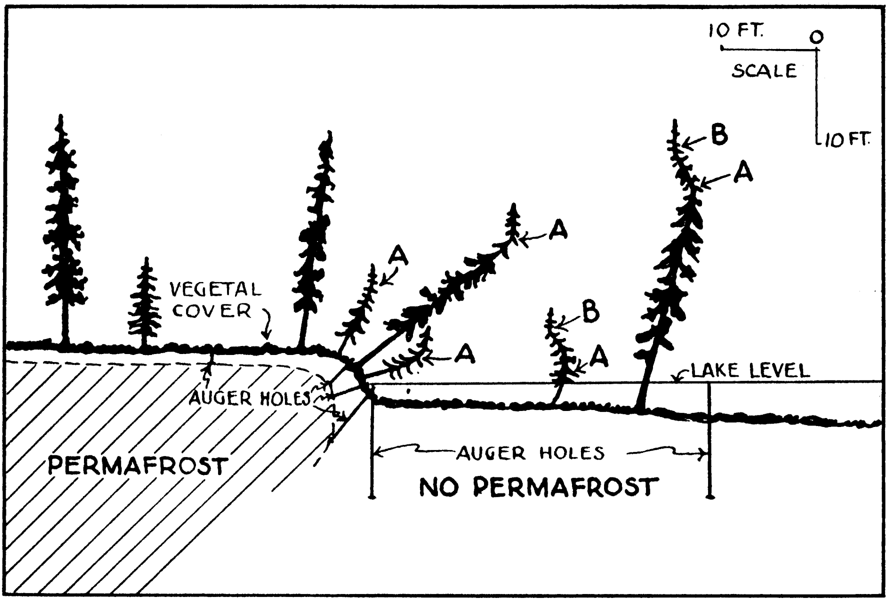



FIG. 3.—Diagram of cross section of bank of cave-in lake, showing tilted and S-shaped trees and distribution of permafrost. *A*, First bend, formed when receding bank reached the base of the tree. *B*, Second bend, formed after receding bank passed beyond the tree.

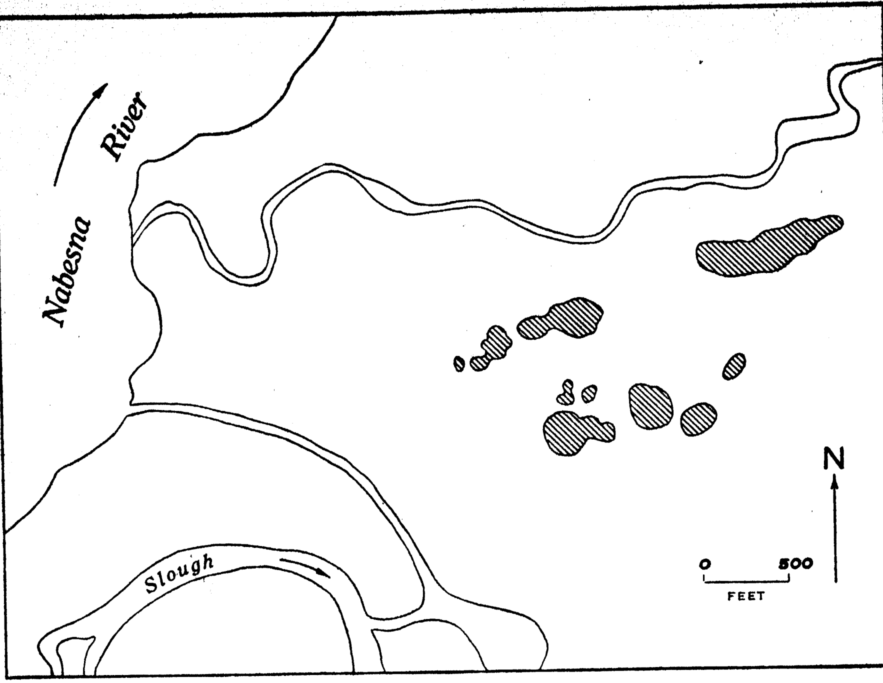



FIG. 4.—Youthful stage of cave-in lake (cross-hatched), marked by simple circular pattern

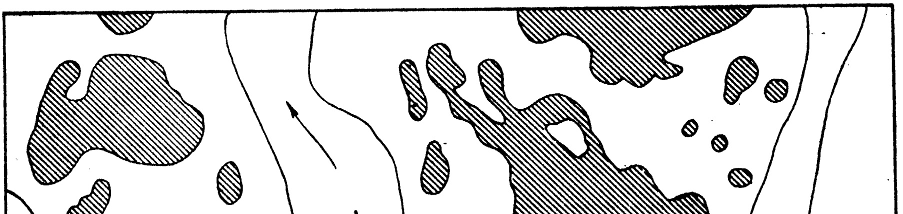



FIG. 5.—Early mature stage of cave-in lake, showing coalescence of many small lakes to form larger lakes (cross-hatched). Note "scalloped" borders. Many youthful, circular cave-in lakes are also present.

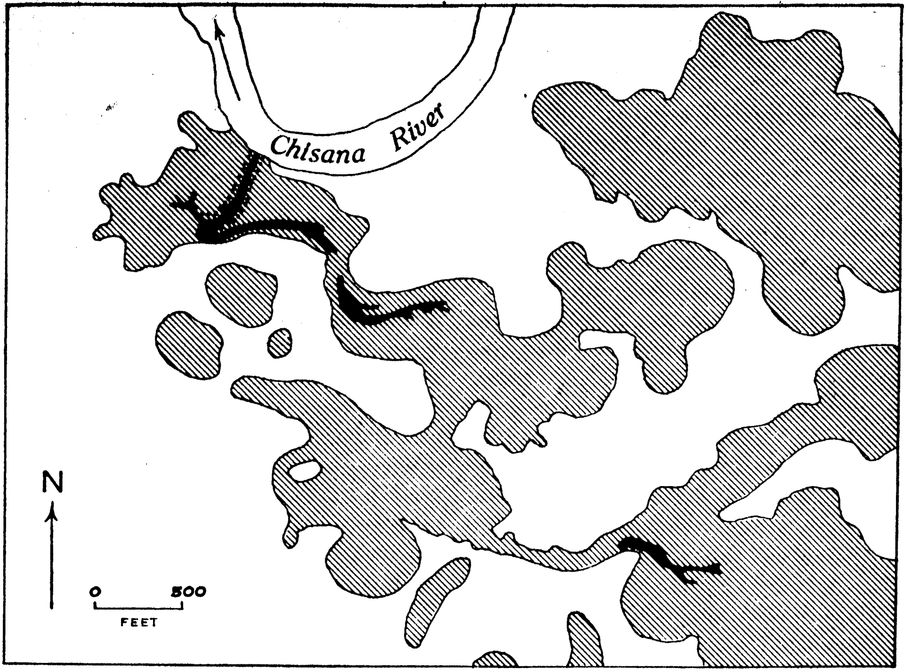



FIG. 6.—Late mature stage of cave-in lake (cross-hatched) showing integration of drainage channels (heavy black) through lake basins, and incipient natural levees (stippled) forming along channels.

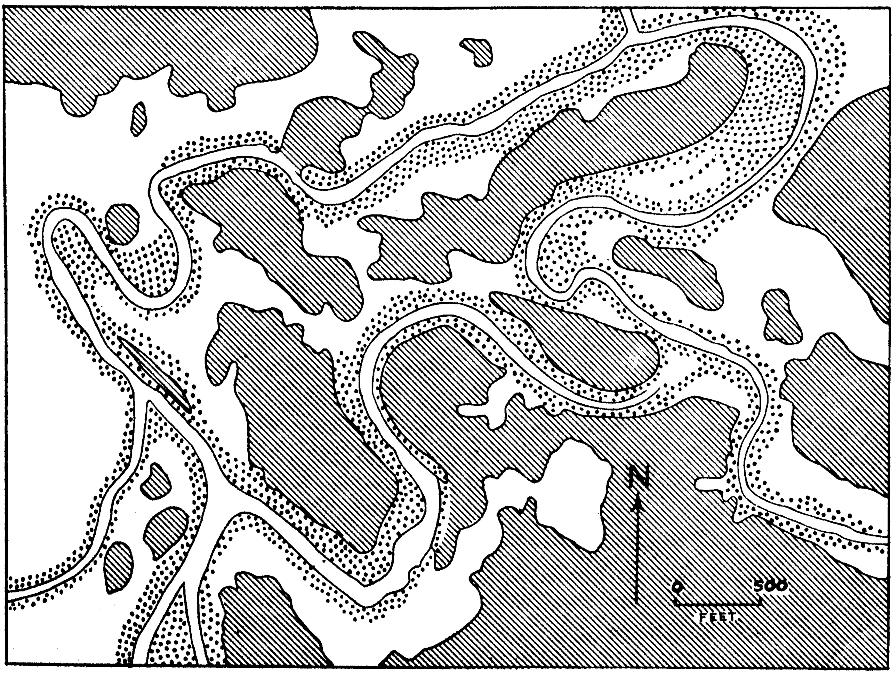



**FIG. 7.—Old age stage of cave-in lake, showing segmentation of lakes (cross-hatched) by natural levees (stippled).**

form, and vegetation takes hold along the borders of the channels crossing the lake basins.

The old-age stage is defined as that in which the development of the natural levees along the drainage channels crossing the cave-in lakes effectively separate the lakes into segments (fig. 7). The levees create the impression that the drainage channels are avoiding the lakes. In some places caving has proceeded up to natural levees already present along streams and has stopped there, probably because of the more stable character of the frozen ground in the levee.

All stages of the cave-in lake sequence occur in the floodplains of the Nabesna and Chisana rivers, and a progression from youthful lakes in the upper parts of the valley to old-age lakes in the lower parts of the valley was noted. The floodplains are almost free of cave-in lakes above a point approximately 9 miles from the mouth of the Nabesna River. Immediately downstream from that point the floodplains are pock-marked with a multitude of small incipient cave-in lakes (fig. 4). Larger lakes and aggregates of lakes representing the early mature stage of development were found at a point approximately 6 or 7 miles from the mouth (fig. 5). Lakes of the late mature (fig. 6) and old-age stages (fig. 7) were found farther downstream, near the junction of the Nabesna and Chisana rivers.

It should be noted that the cave-in lake is the result of a process and that certain features described here are characteristic of only one type of lake. For example, the type of lake described in this paper starts at a point and proceeds outward to form circular borders. In contrast, along the borders of many oxbow lakes, caving has started along a line, and the oxbow lake has been enlarged irregularly. The scalloped characteristic of the lake borders, however, seems to be a typical feature even in such lakes.

## RECESSION OF THE BANKS OF CAVE-IN LAKES

The most direct evidence of the recession of the banks of cave-in lakes is the tilted trees along the borders of the lakes and the drowned trees in the lakes (fig. 3). The rate of retreat of the banks can be determined by noting the progressive changes in the trees affected. A sequence that was commonly observed is diagrammed in figure 3. The undisturbed spruce trees growing on the floodplain stand vertically. As the thawing of permafrost in the banks proceeds, the mat of vegetation, including the shallow, spreading roots of the spruce trees, is undermined and subsides, causing the trees to tilt toward the lake. In the tilted position new growth, growing vertically, forms an angle with the rest of the trunk, as indicated at *A* in figure 3. As the caving progresses beyond the tree, the mat of vegetation and roots again assumes an approximately horizontal attitude but at a lower altitude, so that it is submerged under a foot or two of water. If the roots of the trees are locked in a relatively tight vegetal mat, the tree is tilted toward its originally vertical position. Thus the part that grew while the tree was in a tilted position is tipped from vertical and is tilted toward the bank. Growth continues, and again the new growth, mounting vertically from the tip, forms an angle with the part immediately below, as shown at *B* in figure 3. Such trees can be described as S-shaped because of the two bends in opposite directions.

The first bend, *A* in figure 3, indicates the time in the life of the tree at which it first was tilted. At this time the tree was

at the top of the cave-in bank. The second bend, *B* in figure 3, represents the time at which the base of the cave-in bank reached the tree and the tree was righted.

The number of years that have elapsed since those two events can be determined by counting the internodes from the top of the tree down or by counting the number of growth rings in cross sections of the tree at the bends. The distance that the banks have retreated can be determined and correlated with the elapsed time indicated by the tree growth by measuring the distance from the tree to the base of the bank and to a point on the bank where initial tilting of other trees has begun.

Calculations were made for 15 trees on two lakes. These indicate rates of retreat ranging from 2.3 to 7.5 inches a year. Most of the trees bordering the lakes show but one tilting, that is, toward the lakes. Apparently, the vegetal mat does not remain intact under all the trees, so that, after tilting a certain amount toward the lakes, some trees either fall or at least are not righted after the bank has retreated beyond their position.

The retreating bank is not a simple declivity at all places; indentations and protuberances are common. Small drainage channels develop on the permafrost table along the lake borders at some places. Tributaries branching from these channels were in places nearly parallel to the bank of the lake. Warm meteoric water that flowed in these channels melted the underlying permafrost, and, as it thawed, the trees on either side were tilted toward the channel. Thus some trees on either side were tilted away from the lake and toward the mainland. Such orientations commonly are retained after the trees have been drowned in the lakes. Trees 20 or 30 feet from the banks in some places are tilted in almost all directions.

The vegetal mat containing the roots of trees probably loosens upon saturation with water, so that wind, waves, and heaving of lake-ice, in addition to the thawing of permafrost, can govern the angle and direction of tilt of the trees. Most trees with their roots under water die in less than 10–15 years. Consequently, most of the trees farther out in the lake are dead.

The greatest radius of enlargement that could be identified with relative certainty as having started from a single center is approximately 300 feet. Many radii of approximately this size were noted; but in larger lakes the incorporation of smaller lakes seems to have confused the pattern. At a rate of retreat of 2.3 inches a year, the minimum recorded by trees studied, a lake of 300-foot radius would form in approximately 1,600 years. At a rate of 7.5 inches a year, only 500 years would be required.

### ORIGIN OF CAVE-IN LAKES

The beginning of caving such as that found in the floodplains of the Nabesna, Chisna, and Tanana rivers must be preceded by three events: the deposition of sediments, the volume of which changes with freezing and thawing, followed by development of permafrost in those sediments, and, finally, a change in the thermal balance to cause thawing of permafrost.

How permafrost accumulates or how long it has taken to form can only be inferred by indirect evidence. The presence of permafrost near the surface is directly affected by the insulating ground cover. The coefficient of heat-conductivity of dry peat (Muller, 1945, p. 53) is about 0.06, of wet peat approximately 0.50, and of frozen peat about 2.00. The common

sphagnum moss mat probably has similar characteristics. Thus it is possible for heat to be much more easily transmitted outward from the ground in the winter than into the ground in the summer. The result is an accumulation of cold beneath the moss cover.

Whether or not this process can account for great thicknesses of permafrost under present climatic conditions has not been demonstrated. It may be that older permafrost has migrated upward into the new sediments from the rocks upon which the recent floodplain sediments were deposited. This is known to have taken place in placer-mine tailing piles placed on permafrost-bearing ground. It may be that permafrost has formed almost contemporaneously with the deposition of the sediments and has thus been built up, layer by layer.

The point of lowest stable temperature in one hole drilled to a depth of 44.5 feet in the Nabesna River floodplain by the engineers of the St. Paul District was between 19 and 22 feet below the ground surface; and perceptible annual temperature fluctuations took place to a depth of approximately 15 feet over a test period of one year. The temperature below 22 feet was found to grade gradually to a higher temperature at the bottom of the hole. The position of the minimum stable temperature just below the zone of fluctuating temperature is evidence that heat is flowing upward from depths represented by the lower half of the hole to the point of minimum temperature and that the ground is being cooled from the surface downward.

The lakes do not seem to be formed until after the floodplains are covered by vegetation. It is apparently the change from a good insulator, the vegetal cover, to a good accumulator and distributor of heat, such as the lake waters with relatively high specific heat, that enlarges the lake basins. The effect of removal of the vegetal cover has been demonstrated in many places where caving has taken place in ground that had been cleared for construction purposes. In nature the initiation of a lake could result from such an event as the overturning of a tree by wind, with resulting uprooting of the vegetal mat.

The succession of lake stages, ranging from old age near the mouth of the river to the youthful stage, 9 miles upstream, suggests a correlation with the regimen of the river. In this distance the age of the valley ranges from old age at the mouth to a slightly younger age farther upstream. As baseleveling proceeded, fine-grained sediments, which are suggested as a controlling factor in cave-in lake development, were deposited farther and farther upstream. The earliest deposits of fine material were the first to be covered with a vegetal cover that would insulate the deposits and make possible the formation of permafrost. Thus the conditions necessary for the formation of cave-in lakes would be met first in the oldest part of the valley, and the stages in the lake cycle would range upstream from older to younger in proportion to the degree of adjustment of the river and the vegetal cover on the floodplain.

## RELATED CONSIDERATIONS

A few of many conditions for which conclusions or information were not obtained during this investigation but which warrant study are: first, the possibility that permafrost forms under the cave-in lake basins after they have been drained and a vegetal cover re-formed; second, the possibility that, if permafrost does form under the old cave-in lake basins, a new sequence or cycle of caving is begun; third, the possibility of identi-

fying "fossil" cave-in lake features in areas now south of the zone of perennially frozen ground; fourth, the possibility of recognizing additional physiographic changes due to the widespread formation of permafrost, as, for example, the heaving of ground as contrasted with caving.


SUMMARY

1. Cave-in lakes result from the thawing of permafrost in material which occupies less volume when thawed than it does when frozen. These conditions limit the development of cave-in lakes to the zones of permafrost and principally to situations in which the ground is composed of fine-grained sediments, as in the floodplain deposits of the Nabesna, Chisana, and Tanana rivers. Cave-in lakes are, therefore, a key identifiable in aerial reconnaissance for delimiting ground which slumps when permafrost is thawed.

2. Cave-in lakes possibly form as a result of a break in the insulating ground-cover of vegetation, and their perpetuation and enlargement is a result of the accumulation and distribution of heat by the waters in the cave-in lakes.

3. A sequence of development which can be divided into four stages was recognized. The youthful stage is characterized by circular lakes a few hundred feet in diameter. In the early mature stage the lakes have joined to form aggregates of lakes having "scalloped" shoreline patterns. The late mature stage is reached when an integrated drainage has formed between and through groups of lakes. The old age stage is typified by the predominance of natural levees that have formed along the drainage channels cutting the lakes, thus dividing the lakes into sections. In the old age stage the channels have the appearance of "avoiding" the lakes.

4. S-shaped trees are characteristic of cave-in banks under certain conditions and were used as time gauges in calculating the rate of retreat of banks.

5. The rate of retreat of cave-in banks was found to be between 2.3 and 7.5 inches a year. Ages ranging between 500 and 1,600 years are computed for the lakes in an early mature stage of development.

6. The type of cave-in lake described is one of many manifestations of the cave-in phenomenon.


ACKNOWLEDGMENTS.—Appreciation is expressed to Dr. Siemon W. Muller, of Stanford University and of the Geological Survey, for the introduction to the problem of permafrost. Mr. John Zydik acted as topographer and assistant during this study. Without the cooperation of the Air Transport Command and the Alaskan Department of the Army the execution of the project would hardly have been possible. Excellent aerial photographs of the area made by the U.S. Army Air Force were most useful.

REFERENCES CITED


BRYAN, K. (1946) Cryopedology—the study of frozen ground and intensive frost-action with suggestions for nomenclature: Am. Jour. Sci., vol. 244.

MULLER, S. W. (1945) Permafrost or permanently frozen ground and related engineering problems: Office, Chief of Engineers, Special Rept. Strategic Eng. Study 62.

SMITH, P. S. (1939) Areal geology of Alaska: U.S. Geol. Survey Prof. Paper 192.

TABER, S. (1929) Frost heaving: Jour. Geology, vol. 37, pp. 428–461.

# CRATERS AND CRATER SPRINGS OF THE RIO SALADO[1]

E. R. HARRINGTON
Albuquerque, New Mexico

ABSTRACT

The craters and the crater springs are located approximately 8 miles, air line, northwest of the Jemez Indian Pueblo, which is some 45 miles northwest of Albuquerque, New Mexico. They lie on the eastern slope of the valley of the Rio Salado, a tributary of the Rio Jemez, which, in turn, flows into the Rio Grande. The springs rise through faults in the Permian, Triassic, Jurassic, and Cretaceous sediments, which dip steeply to the west from the Nacimiento uplift first described by Herrick (1904). The area is part of the "Ojo del Espiritu Santo" grant, and the springs here mentioned are referred to locally as the "Phillips springs."

Travertine crater springs are not especially unusual. The craters of the Rio Salado, however, are on a very grand scale and show some features that may be unique. The spring area offers unusual opportunity to study such formations. Within an area of 1 square mile there are some forty springs or craters, which show all gradations from incipient formation to complete extinction.

Probably North America offers no better area for the study of travertine springs. It is surprising that the area is so little known to the continent's geologists.

The region was first visited by white men in 1540, when Coronado, the Spanish conqueror, entered the area in search of gold. The springs of the Jemez country, a few miles to the east, were mentioned in the writings of the time; and it was known that the Coronado expedition mined sulphur for gunpowder a few miles to the east. A few years later some other Spanish explorers saw the great crater springs and named one of them "El Ojo del Espiritu Santo," the "Spring of the Holy Spirit." The name has come down to the present time, indicating that the explorers of the early day were definitely impressed by what they saw. In later times the Spanish colonization of the whole watershed of the Rio Salado was administered as one grant from the Spanish crown, and today it still carries the name of the "Ojo del Espiritu Santo" grant on all maps and legal documents.

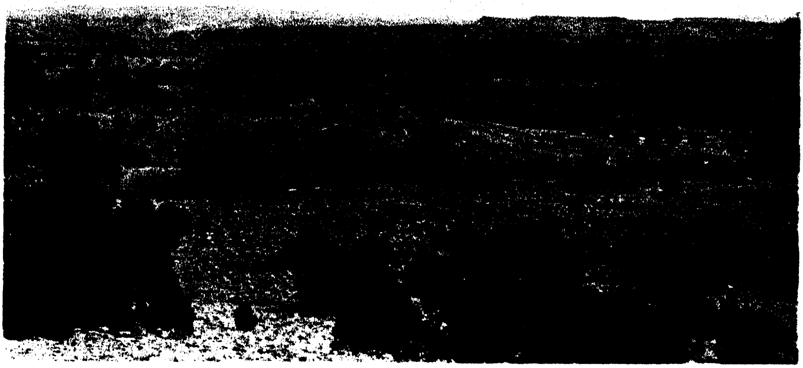
Following the Mexican war in 1845, the area passed to the ownership of the United States. A year later Abert (1848*a*, pp. 3–130, map; 1848*b*, pp. 417–546, map) made a geological examination of the newly acquired territories and, when traversing the Jemez region, noted the large deposits of travertine. No references were made specifically to the crater spring area until after 1900, when the waters of the Jemez plateau were studied by Kelly (1913, p. 76). Later, in 1929, Clark (1929, p. 30) made a study of the saline springs of the Rio Salado; and two years later Renick (1931, pp. 87–88) carried on a ground-water study of the same area. In 1943 Harrington (1943, pp. 7–12) referred specifically to the crater springs.

The springs and craters all lie west of the Nacimiento uplift. They rise from the Permian "red beds," which dip steeply away from the pre-Cambrian core of the Sierra Nacimiento. The older craters were high up on the sedimentary slope. The spring waters welled to the surface

[1] Manuscript received October 12, 1947.

A

*A*, Large travertine cone. This one rises more than 300 feet above its base. It contains a dry crater, 100 feet in diameter and more than 100 feet deep. It was drained by another crater which broke out several hundred feet below it.



B

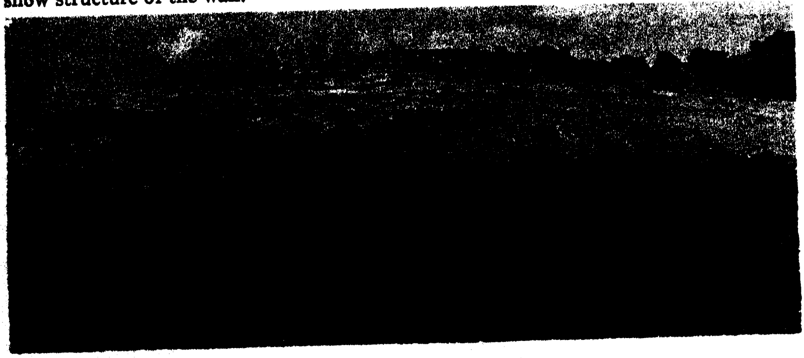
*B*, Looking down on the crater which drained the crater shown in the preceding picture. This crater had built itself up to a height of more than 100 feet. This crater flowed until 1926, when it was partially drained by a flowing well drilled a mile away. Vegetation has not yet had time to make an effective attack on this crater, and it has retained its "bald" appearance.

C

*C*, Close-up of the crater of the preceding photograph. Crater is about 25 feet across and has vertical sides going down at least 80 feet. The water inside the crater now is about 30 feet below the rim.

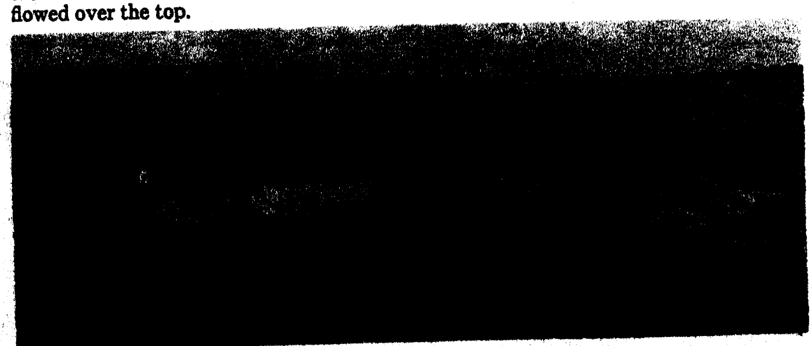
A

*A*, View looking down into the crater of photograph *C*, above. Picture was taken to show structure of the wall.



B

*B*, Large travertine crater. This cone is at least 200 feet high. The left slope near the bottom shows an active vent which is now draining away the water which once flowed over the top.



C

*C*, A view of the "Swimming Pool" spring. Crater is more than 60 feet across and 70 feet deep. Crater is nearly circular. Flow from spring is about 10 gallons per minute.

through a fault. All the waters were highly charged with carbon dioxide, salt, and gypsum. At the surface the loss of carbon dioxide caused part of the calcium bicarbonate to be precipitated as calcium carbonate, and travertine rims were built around the spring vents. The water, of course, ran across the lowest part of the rim; and, as it did so, that part of the rim was built up, and the water was forced to seek a new channel across the confining travertine edge. After centuries of such deposition the springs grew into great travertine cones or mounds, with large craters in their centers (pl. 1, *A*). Some of these craters, now completely dry, consist of a great mound of travertine more than 400 feet high. In one of the large extinct craters the central vent has a diameter of more than 100 feet and a

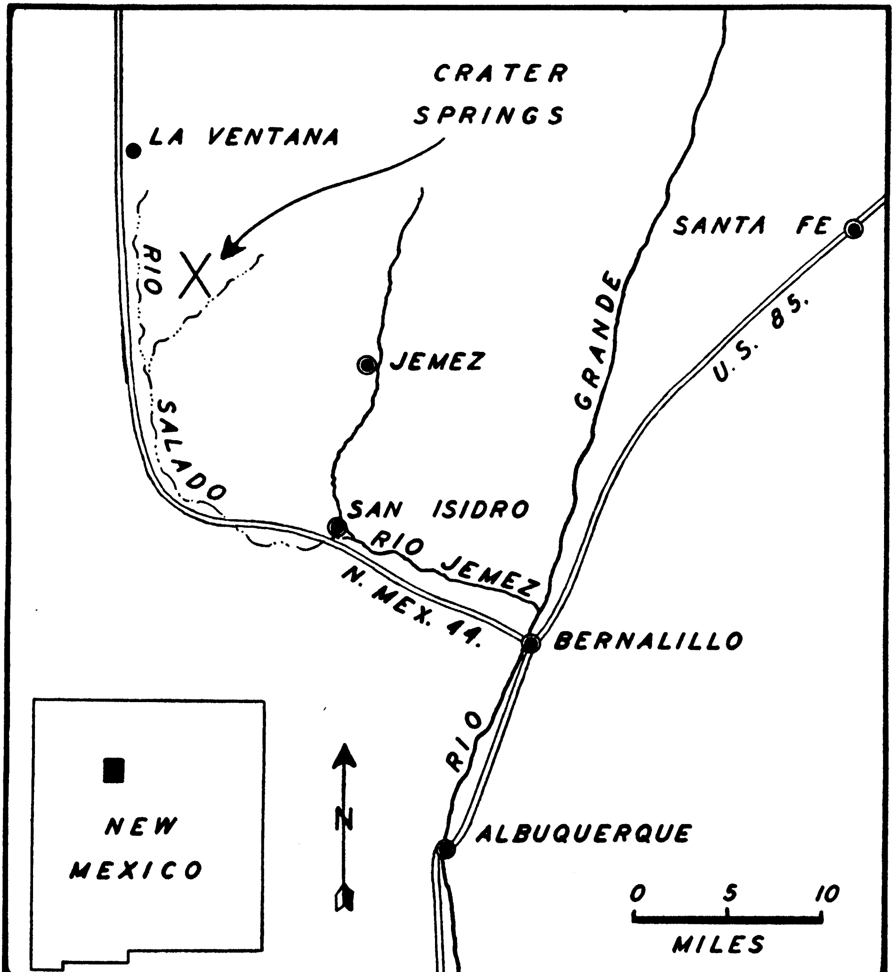



FIG. 1.—Map of the Rio Salado area of New Mexico

depth of possibly 150 feet. After the great craters had built themselves up to such a height, the hydrostatic pressure at the bottom became considerable, and the water broke out farther down the slope. When this happened, the original crater was drained, and a new crater formed at a lower level. In one case three craters show three stages of this drainage process. The top crater was drained by the second one, and the second one was drained by a third. The third one was still building up in 1920. A few years later a well was drilled for gas a mile away. This well struck hot artesian water and has been flowing ever since it was drilled in 1926. The flowing well was several hundred feet lower than the third crater of the series just mentioned. The relief of pressure caused the third crater to cease flowing, and the water in that crater now stands at about 30 feet below the crater rim (see pl. 1, *B* and *C;* pl. 2, *A*). The water in the third and lowest crater was at a temperature of about 70° F., while the water from the hot well issued at about 125° F. From the difference in temperature it appears that there was no common source of water for both well and crater; but it is also evident that some sort of fissure connection must exist between the two sources because the drilling of the well caused the crater to cease flowing. This recently active crater has an irregular vent some 25 feet across. The vent has perpendicular sides and a depth of about 80 feet, the bottom 50 feet being filled with water (see pl. 2, *A*).

Half a mile south of the craters mentioned in the preceding paragraph, the Penasco Creek cuts a channel into the Rio Salado. Along the banks of this creek there are several fine examples of the crater springs. One is a dry crater, 20 feet across and 50 feet deep. Above this crater is a "living" crater spring which has built up a cone more than 100 feet high. Several gallons of water per minute flow from this spring, which is extremely salty and heavily charged with carbon dioxide. The crater vent is only about 10 feet across, but a sounding line was sent down more than 50 feet before encountering any obstruction. Despite the extreme salinity of the water, some hardy salt grasses grow around the crater vent. A quarter of a mile farther up the canyon of the Penasco Creek there is another large travertine crater built to a height of possibly 200 feet (pl. 2, *B*). This crater has twin vents, which are small and not unlike those of geysers. In one of these vents the water stands at about 15 feet below the top. Some 50 feet below, on the east side of the crater, a new vent has formed, and this outlet is draining the water from the vent above.

But half a mile farther up the Penasco Creek is the largest of the still active springs. The flow is less than 10 gallons per minute. Analysis shows that the water from this spring carries more than seven thousand parts per million of solids. The water is salty and heavily charged with carbon dioxide, large bubbles of this gas "boiling up" in the center of the crater lake. This spring crater is an almost perfect circle, with a diameter of more than 60 feet. A sounding line showed the sides of the crater to be almost vertical and the depth to be in excess of 70 feet. Algae growing in the water give it a dull gray-green color sufficient to make a foot of the water opaque. The water temperature is about 70° F. This crater spring is called the "Swimming Pool" (pl. 2, *C*). The name is derived from its general appearance rather than because of its desirability as a place to swim (the writer did swim in it, but he does not recommend the procedure as being safe).

I have seen many travertine deposits and have spent some time in study in Yellowstone Park. I believe that the crater springs of the Rio Salado show phenomena that are not surpassed elsewhere. The great craters are only a mile or two away from a paved road and less than 50 miles from New Mexico's largest city. Even so, the craters are not well known and are seldom visited. This should not be so, and I wish to call the attention of scientists to the opportunities for study presented by the crater springs of the Rio Salado.

## REFERENCES CITED

ABERT, J. W. (1848*a*) Report of the examination of New Mexico in the years 1846–47: U.S. 30th Cong., 1st session, Senate Executive Document 23, pp. 3–130, map.

——— (1848*b*) Report of the examination of New Mexico in the years 1846–47: U.S. 30th Cong., 1st session, House Executive Document 41, pp. 417–546, map.

CLARK, J. D. (1929) The saline springs of the Rio Salado: Univ. New Mexico Bull. 163.

HARRINGTON, E. R. (1943) Springs of the Jemez: New Mexico Mag., June, pp. 7–12.

HERRICK, H. N. (1904) Gypsum deposits in New Mexico: U.S. Geol. Survey Bull. 223, pp. 89–99.

KELLY, CLYDE, and ANSPACH, E. V. (1913) A preliminary study of the waters of the Jemez Plateau: Univ. New Mexico Bull. 71.

RENICK, B. C. (1931) Geology and ground water resources of western Sandoval County, New Mexico: U.S. Geol. Survey Water Supply Paper 620.

# A PRELIMINARY REPORT ON VERTEBRATES FROM THE PERMIAN VALE FORMATION OF TEXAS[1]

EVERETT CLAIRE OLSON
University of Chicago

ABSTRACT

Field work during 1946 and 1947 has produced a moderately extensive fauna of fish, amphibians, and reptiles from the Vale formation of Knox County, Texas. The youngest terrestrial vertebrates previously known from the Texas Permian are of Arroyo age. Aerial photographs were used successfully in exploration and as a basis for accurate indexing of localities.

The vertebrates occur exclusively in channel conglomerates laid down in moderately evenly bedded red and green sands and clays. The environment appears to have been one of moderate rainfall, with extended dry periods interrupted by torrential rains.

*Dimetrodon* and *Diplocaulus* occur with the greatest frequency. Xenacanth sharks, dipnoans, gymnarthrids, lysorophids, diplocaulids, captorhinids, and various pelycosaurs make up much of the fauna. A preliminary survey indicates that the Vale fauna was derived from the Arroyo fauna, with the addition of a few new elements. A well-developed flora is preserved but is as yet unstudied.

## INTRODUCTION

During the summer of 1946 a field party under the writer's direction located vertebrate-bearing beds in the upper part of the Vale formation of the Clear Fork Permian, in Knox County, Texas. Extensive collections of vertebrates have been obtained from the Clyde and Arroyo formations of the Clear Fork group during a period of some seventy years, but all attempts to find terrestrial vertebrates in the Clear Fork formations higher than the Arroyo had been unsuccessful prior to 1946. Aerial photographs, used for preparation of detailed maps (fig. 1) and for reconstruction of the paleogeography, plus a measure of good fortune, made the new discoveries possible. Photographs aided in suggesting that exploration of channel deposits would be most profitable and in plotting routes through the rough country; chance prompted the establishment of the base camp on the only one of several equally inviting localities which has as yet shown any potentiality for production.

A small quarry, which yielded several specimens, was opened in 1946. In the spring of 1947 adjacent deposits were studied. At present, producing beds are known to occur in a north-south band about 1 mile wide and about 8 miles long (fig. 2). A moderately extensive fauna, consisting of fish, amphibians, and reptiles, has been obtained. The fauna has the general cast of that from the Arroyo but differs from it in a number of respects. Associated is a rather extensive flora, as yet unstudied, which should prove to be of considerable interest.

## LOCATION OF THE DEPOSITS

The vertebrate-yielding beds of Vale age extend northward about 8 miles from an area located along the southern margin of the breaks of the South Wichita River, 4 miles north of Vera, Texas, on the Vera-Gilliland road in Knox County, Texas. The northernmost exposures yet visited lie along the southern margin of the breaks of the North Wichita River. Between these two areas occur exposures in scattered patches in the valley of the South Wichita River and along the northern margin of its valley (fig. 2).

The productive outcrops are shown on

[1] Manuscript received November 15, 1948.

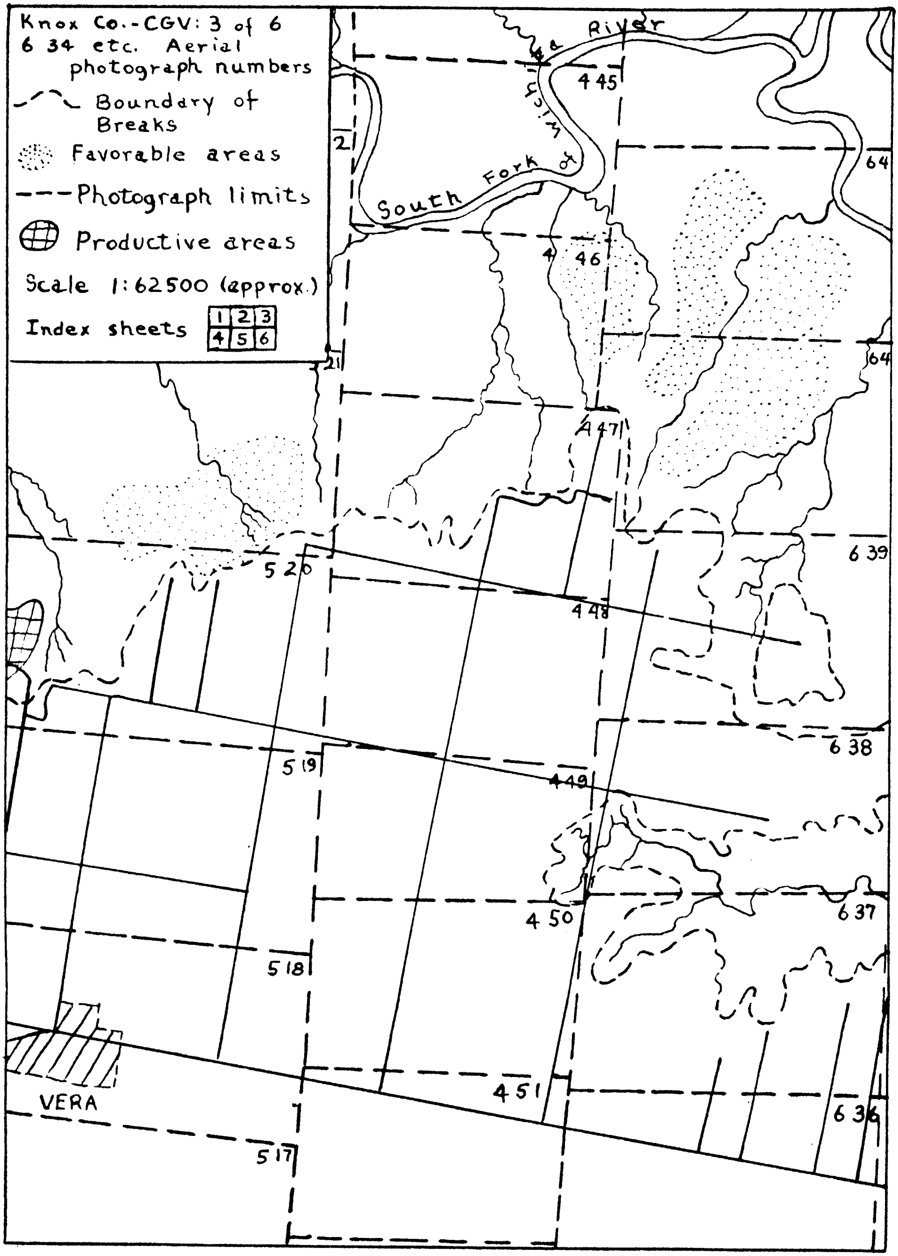



Fig. 1.—An example of the type of map prepared from aerial photographs for field reconnaissance. Reduced from original scale of approximately 1:62,500.

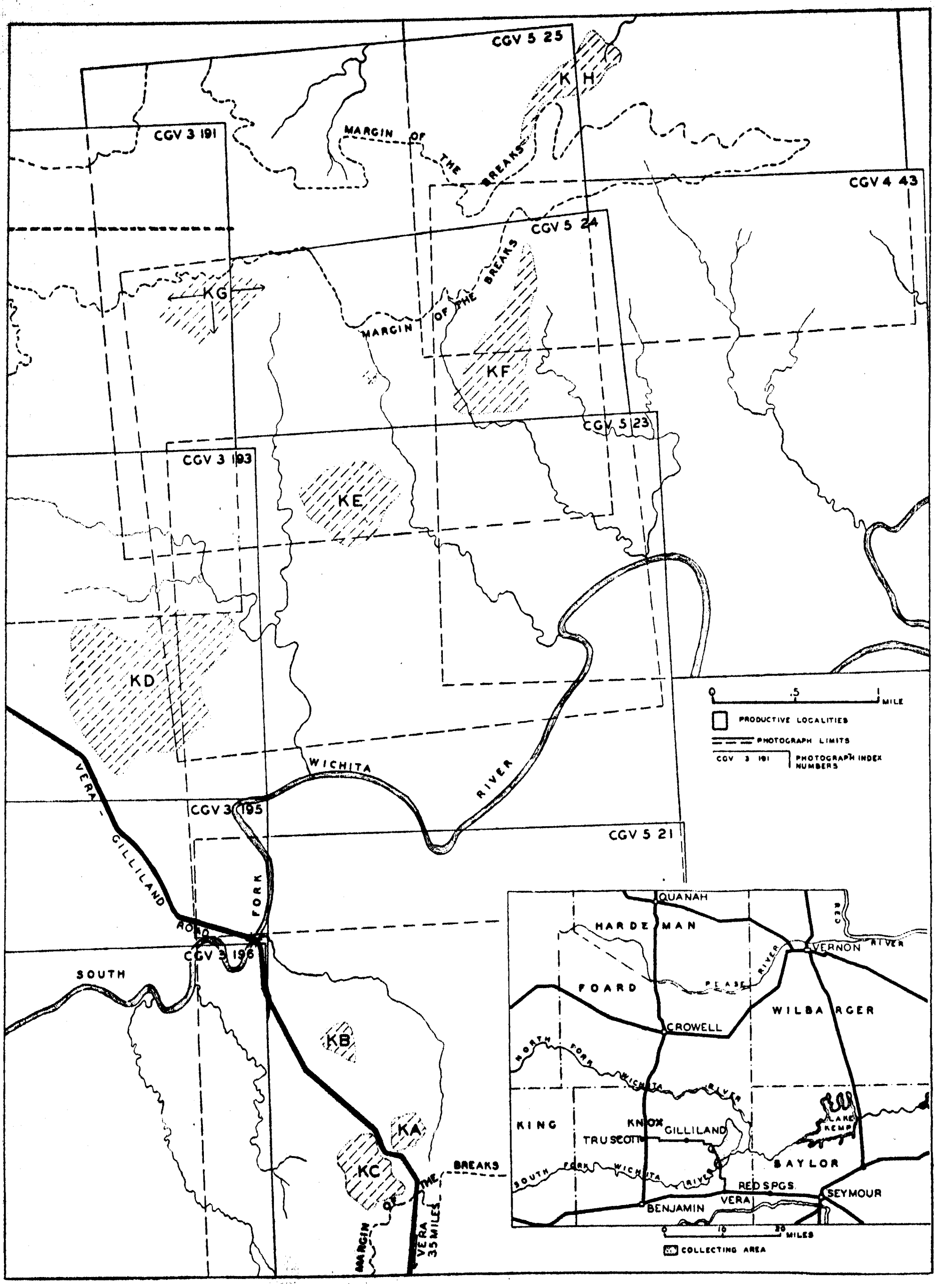



FIG. 2.—Map of collecting area in Knox County, north central Texas, based on aerial photographs. *KA*, *KB*, etc. indicate localities from which fossils have been obtained. Inset map shows the position of collecting area relative to adjacent counties.

aerial photographs of the United States Department of Agriculture of Knox County, *CGV 5 21–25* and *CGV 3 191–194*, as keyed in figure 1. For purposes of reference, each locality is designated by two letters. The first letter, *K*, indicates location in Knox County, to distinguish localities from those in other counties. The second letter distinguishes each locality from others in Knox County. The second letters have been entered in the order in which the localities have been discovered and have no other significance. Each lettered locality may be located precisely by the use of photocoordinates. In the system employed, the locality is defined by three or four limiting points, and, if needed, a roughly central point is also designated. Spot localities of important sites are also noted. All co-ordinates are measured in inches from the lower left-hand corner of the photograph with the *X*-axis entered first and the *Y*-axis second. Small localities are expressed to $\frac{1}{100}$ of an inch which, on 1:20,000 photographs, locates a point within a circle of about a 17-foot radius. Locations of productive areas visited to date are as follows:

KA South wall of valley of South Wichita River east of Vera-Gilliland road

Photograph.............. CGV 5 20
Centered at.............. 3.2–6.9
Individual localities........ 3.2–7.1*; 3.3–6.85†; 3.2–6.77‡

* "Bone bed."
† "High channel."
‡ "West channel."

KB Valley of South Wichita River, south of river and east of Vera-Gilliland road

Photograph.............. CGV 5 21
Centered at.............. 2.55–4.9
Triangular area limited by.. 2.1–5.1; 2.5–5.1; 2.6–4.9

KC South walls of valley of South Wichita River, west of Vera-Gilliland road

Photograph.............. CGV 5 21
Roughly rectangular area limited by.............. 3.7–3.8; 3.1–3.2; 2.5–2.7; 2.2–3.2

KD Valley of South Wichita River, north of river and east of Vera-Gilliland road

Photograph.............. CGV 3 193
Centered at.............. 6.8–4.8
Bounded by irregular lines following outlines of breaks between.............. 8.3–3.5; 7.3–3.5; 5.5–5.8; 6.3–3.1
Individual localities........ 6.4–5.2§; 7.18–4.22‖

§ "Quarry locality."
‖ "Weathered-boulder locality."

KE Valley of South Wichita River, north of river and east of Vera-Gilliland road

Photograph.............. CGV 5 24
Centered at.............. 3.5–4.1
Roughly rectangular area limited by.............. 4.5–4.3; 3.6–4.8; 2.7–4.3; 3.6–3.4

KF North wall of valley of South Wichita River, east of Vera-Gilliland road

Photograph.............. CGV 5 24
Roughly triangular area limited by................ 7.4–8.4; 5.6–5.7; 7.0–5.6

KG North wall of valley of South Wichita River, east of Vera-Gilliland road

Photograph.............. CGV 5 25
Centered at.............. 2.0–4.8
Precise limits not as yet defined

KH Southern margin of valley of North Wichita River

Photograph.............. CGV 4 41
Strip ½ mile wide from..... 1.9–3.7 to 3.2–4.6
Limits not fully defined as yet

This type of designation has proved invaluable in the field. Each locality, of

course, appears on more than one photograph; the practice is to cite the photograph upon which the exposures are best shown. The system has been applied in Baylor County as well, and, as previously known localities are visited, photo-index designations are given. Some twenty localities have now been so designated. With this system, "lost localities" should become a thing of the past, and data for stratigraphic work will be much more exact.

## GEOLOGY

### AGE OF THE BEDS

There is little in the immediate vicinity of the vertebrate-bearing strata which has any obvious bearing upon their exact age. The Lueders limestone crops out about 20 miles to the east, and upper Clear Fork marine beds cap the hills in western Knox County. Between these limestones no definitely marine beds are encountered. As yet, neither the vertebrates nor the plants have proved definitive within the limits needed, and no invertebrates have been found. Robert Roth, of the Humble Oil and Refining Company of Wichita Falls, Texas, believes that surface and subsurface data show that the beds in question are part of the "Bullwagon" member of the Vale formation. This places the section high in the Vale, just below the Choza.

All fossils have come from rocks which were deposits in channels cut in rather evenly bedded clays and sandstones. The channel deposits are, however, essentially contemporaneous with the beds in which they lie, as discussed in following sections.

### NATURE OF THE BEDS

Fossil-bearing beds are exposed along the steep walls of the South Wichita River Valley, in scattered outcrops on the floor of the valley, and along the south wall of the valley of the North Wichita River. Exploration has not been carried across the valley of the North Wichita, but it appears probable that fossiliferous beds exist along the northern margin of this river. Permian beds form the floor of the South Wichita Valley, but along the margins they are capped by indurated Pleistocene sands and gravels with thicknesses ranging up to 30 feet.

The Permian beds may be divided roughly into two types: (1) widespread, rather evenly bedded red and green clays, sands, and fine conglomerates and (2) linear channel conglomerates.

Exposures of clays, sands, and fine-bedded conglomerates extend for several miles up and down the valley of the South Wichita River. To the east they gradually change in character to grade into typical Arroyo beds around Lake Kemp in Baylor County. Their western extension has not been studied; but it is known that there is an increase in the regularity of the bedding, a reduction in the amount of conglomerate, and gradation into the brackish and marine phases of the Choza, in which primary layers of anhydrite and limestone are well developed. The strata have been studied in some detail for about a mile to the east and west of the vertebrate-bearing beds. Within this distance they are relatively constant and, except in the immediate vicinity of the channel conglomerates, are moderately evenly bedded and include members with considerable lateral continuity. As the channel zone is approached, bedding becomes less regular, and dips up to about 12° are encountered. Gypsum is present in irregular masses in the bedded clays and sands and in dikes. A few thin, horizontal layers of fibrous gypsum, with fibers at right angles to the bedding, have been encoun-

PLATE 1

Two views of outcrop of channel deposits at "bone bed" of locality *KA*. *Upper view:* Full extent of outcrop, quarry under ledge in left foreground. *Lower view:* Margin of channel showing rounded lateral edge of conglomerate.

PLATE 2

Section cut through typical sample of red, channel-fill conglomerate

tered. Most, if not all, of the gypsum appears to be secondary; its source apparently was anhydrite beds not now exposed in the area. Primary beds are exposed farther to the west, in the breaks north of Benjamin, for example.

It is the beds to the west which must be explored if vertebrates from still higher Permian formations are to be found in the region. The fact that these are evenly bedded and rich in anhydrite, coupled with the fact that the evenly bedded rocks in the locality under consideration have failed to produce vertebrates, makes prospects rather poor. It seems fairly certain that only channel deposits will be fossiliferous. Reconnaissance studies to date have failed to locate channels in the higher beds.

The channel deposits of the area under discussion in this report, although limited in extent, are of prime interest, since they contain all the vertebrate specimens thus far discovered. The channels are randomly spaced throughout a 100-foot section. The lowest exposures occur on the valley floor in tributaries of the main channel of the South Wichita River not more than 10 feet above river level, and the highest occur directly under the Pleistocene. Several levels of channels have been witnessed, one above the other, in a single exposure. No case in which one channel cuts another has been observed. There appears to be no real faunal difference between the lowest and the highest channels, and, since indications of rapid sedimentation are abundant, age differences appear to be insignificant.

All channel deposits are conglomeratic and distinct from the adjacent beds. They tend to be more resistant to weathering and erosion than are the clays and sands, and they form escarpments and falls and rapids in the intermittent tributaries of the main river (pl. 1). The deposits are distinctly linear, and individual channels have been traced for as much as a mile along meandering courses. Although individual channels are randomly oriented, the general trend of the system of channels is north-south.

The channel-fill is largely clay-pebble conglomerate, with pebbles ranging from a few millimeters to about 8 cm. in diameter (pl. 2). Rounding of the pebbles is uniformly high, but sphericity is varied. There is little vertical variation except near the top of a few channels. The coarsest materials tend to lie at the center of the channel, whereas the marginal parts are predominantly sand or clay with fine, scattered pebbles. Some channels have, in addition to the clay pebbles, fragments of gypsum. Some of these appear to have been developed in openings in the porous conglomerate after deposition of the channel-fills, but others show rounding and wear, which suggest that they are clastic.

In cross section the channels are characteristically lenticular, with convex surfaces above and below. The lateral margins terminate rather abruptly in clearly defined edges, which show no continuity with adjacent beds (pl. 1, lower, and pl. 2). This pattern holds for all major channels and for the majority of small ones; but a few, particularly those which form lenses in sandstone, are less regular and grade laterally into the even beds. The bases of the channel deposits do not grade into the underlying beds, and the tops are commonly distinct. In a few instances, however, the channel conglomerate grades upward into a sandstone which is continuous with the lateral beds.

The coloration of the channels appears to have significance. Channels composed entirely of red or of green sediment have been encountered; much more com-

mon, however, are those in which the center portion is red and the lateral parts are green. The transition is accomplished by an increase in green patches, which occur in some cases even in the center of the channels, until they become dominant and at the extreme margins replace the red entirely. The explanation of color variation is not clear as yet. It is possible that the transition from red to green is due to greater reduction of the lateral portions of the channels, if it is assumed that the primary color was red. On the other hand, the original deposits may have been green, and the red central portions could be due to oxidation of the coarser center parts of the channels. The evidence at hand from preliminary studies suggests that the first interpretation is correct, but the data are by no means conclusive. In channels which are partly red and partly green the original deposition cannot account for the arrangement of materials of the two colors, so that a postdepositional factor clearly was in operation. Megascopic plant remains occur in both the red and the green phases of the channels and have not accomplished notable reduction in the portions which have remained red. Bacteria may have been an important factor, but there is no positive evidence that this was the case.

The relation of the color to the vertebrates is most important. With one exception, all specimens have come from the red phases of the channels. No specimens have been found in the completely green channels, but some bone was encountered in the fine, green lateral phase of a channel in locality *KF*. Preservation of these specimens is good, and whatever changes may have taken place to produce the green color appear to have had no important effect on the bones. That the absence of bones in the green sediments is due to the reduction which altered the red deposits to green is an inviting hypothesis and, except for the one occurrence of bone in green sediment, is in accord with the known facts.

### ENVIRONMENT OF DEPOSITION

A reconstruction of conditions at the time of deposition appears to be possible on the basis of the information now at hand.

The channels are evidence that a rather large, braided stream passed north or south over the area in a valley of considerable width. The region must have had a moderately high annual rainfall to support the fauna and flora which have been encountered. The total rainfall was perhaps not greatly different from what it is today, and the temperature may have been quite similar, with possibly less seasonal variation. The rains, however, appear to have been periodic and torrential in character, for it is difficult to account for the deposition of the clay-pebble conglomerates in channels under any other interpretation. During a dry period it would seem that little or no water flowed through the area and that silting in the final stages of flood and shifting of dry sediment by wind action obliterated much of the drainage pattern developed in the preceding rainy period. Were this not the case, it would be reasonable to suppose that earlier channel deposits would be cut by subsequent channels, but no case in which this has occurred has been noted.

During the dry periods aquatic life must have been limited to whatever rain pools, ponds, or small lakes may have been present. The lungfish may have survived by burying themselves in the mud, much as they do today; and there is no proof that the amphibians may not have

behaved similarly. The principal amphibians are *Diplocaulus* and *Lysorophus*, one apparently rather skatelike in shape and the other wormlike. With the onset of heavy rains, floods appear to have gouged out channels which bore little or no relationship to former channels and, with rapid drop in velocity as the rains ceased, to have deposited clay, cut from their banks, as pebbles in a fine matrix of sand and clay. During late stages of flood, as the current slowed and the water began to recede, even beds of sand and clay were deposited on the floodplains and apparently over the channels themselves. This cycle of aggradation appears to have been repeated many times during the deposition of the beds under consideration.

### THE FAUNA

Preparation of the specimens collected in 1946 and 1947 is proceeding slowly; and, since it will be some time before a comprehensive report can be issued, the following preliminary account is felt to be in order. Most generic references must be considered tentative, except where assignments are stated to be definite. Names are witheld from genera which are almost certainly new, pending development of all available material. It is hoped that additional collecting will produce more specimens of problematic genera. All known specimens, with one possible exception, appear to be referable to orders recorded from the Arroyo formation, and in most instances specimens can be assigned to families represented in this formation.

### PRESERVATION OF THE SPECIMENS

Almost all specimens have come from coarse channel conglomerates. As might be expected under such circumstances, most of the material is fragmentary. An articulated series of vertebrae of *Dimetrodon* was recovered from the quarry of locality *KA;* and partial vertebral columns of small animals, with ribs and parts of the appendicular skeleton associated, have been found in several other localities. Partial skulls, lower jaws, isolated teeth, and various postcranial elements make up the bulk of the collections. Disarticulation of the skeletons has made association of parts of new genera difficult, and the situation may not be entirely clarified until more complete specimens are discovered.

The bone itself is normally in an excellent state of preservation, and, in spite of the coarse matrix, delicate structures have remained intact. The process of preparation has been complicated by the soft and brittle or waxy nature of the bone. Once the matrix has been removed, however, detailed analyses of morphology should prove possible.

### CONSTITUTION AND DISTRIBUTION

A tentative general faunal list of identifiable specimens collected to date from the Vale formation of Knox County is as follows:

- Class Chondrichthyes
  - Order Xenacanthodii
    - Family Xenacanthidae
      - *Xenacanthus* sp.
- Class Osteichthyes
  - Subclass Choanichthyes
    - Order Dipnoi
      - Genus undetermined
- Class Amphibia
  - Subclass Lepospondyli
    - Order Nectridia
      - Family Diplocaulidae
        - *Diplocaulus* sp.
    - Order Micramphibian
      - Family Gymnarthridae
        - Genus undetermined
    - Order Lysorophia
      - Family Lysorophidae
        - *Lysorophus* sp.

Class Reptilia
- Subclass Eureptilia
  - Infraclass Captorhina
    - Order Captorhinomorpha
      - Family Captorhinidae
        - *Captorhinus* sp.
        - ?*Labidosaurus* sp.
  - Infraclass Synapsida
    - Order Pelycosauria
      - Suborder Sphenacodontia
        - Family Sphenacodontidae
          - *Dimetrodon* sp.
      - Suborder Edaphosauria
        - Family Edaphodauridae
          - ?*Edaphosaurus* sp.
          - Genus undetermined

There is no real basis for specific differentiation from the Arroyo specimens, but teeth of *Xenacanthus* are highly unsatisfactory for specific determination. Several hundred teeth have been observed in the channel conglomerates.

### Class Osteichthyes

#### Order Dipnoi

Several fragmentary specimens of lungfish teeth have been observed, and two moderately good specimens have been obtained. The occurrences are limited to localities *KD* and *KF*, but this probably has little real significance in view of the small size and fragmentary nature of the specimens. The teeth resemble those of *Gnathorhiza* from the Arroyo but show differences in form and size, which make it probable that they must be referred to a new genus.

TABLE 1

| Genera | Localities | | | | | | | |
|---|---|---|---|---|---|---|---|---|
| | *KA* | *KB* | *KC* | *KD* | *KE* | *KF* | *KG* | *KH** |
| *Xenacanthus* | × | × | × | × | × | × | × | × |
| Dipnoan | | | | × | | × | | |
| *Diplocaulus* | × | × | × | × | × | × | | × |
| Gymnarthrid | | | | × | | × | | |
| *Lysorophus* | | | | × | | × | | |
| *Captorhinus* | × | × | | ?× | | | | |
| ?*Labidosaurus* | | | | × | | | | |
| *Dimetrodon* | × | × | ?× | × | × | × | ?× | |
| ?*Edaphosaurus* | × | | | × | | ?× | | |
| Edaphosaurid | | × | | × | | | | × |

* Not fully prospected.

Ten distinct genera are recorded in this faunal list. In addition, there are specimens of amphibians and reptiles which cannot be given family reference as yet and which are distinct from the listed genera. The recorded genera are distributed throughout the various collecting localities, as shown in table 1.

## DISCUSSION OF THE FAUNA

### Class Chondrichthyes

#### Family Xenacanthidae

This group of fresh-water sharks is represented throughout the area by teeth and fragments of calcified cartilage. The teeth are uniformly small and resemble those in the Arroyo formation.

### Class Amphibia

#### Subclass Lepospondyli

#### Order Nectridia

#### Family Diplocaulidae

*Diplocaulus* is the most abundant tetrapod in the area and has a frequency almost double that of its nearest rival, *Dimetrodon*. There is no doubt concern-

ing the identity of the genus. Several good skulls, parts of skulls, and fragments of postcrania have been obtained. At present it is not feasible to attempt specific determination because of the confusion which exists concerning the Arroyo diplocaulids. The writer is now engaged in an analysis of the growth and variations of the Arroyo specimens of *Diplocaulus;* and, when this study has been completed, it may be possible to place the Vale diplocaulids accurately. At present it is impossible to determine whether or not more than one species is present in the Vale.

The skulls are of moderate size, with a midline length of from 80 to 100 mm. They exhibit mature growth patterns, in contrast to skulls of similar length from the Arroyo. The dermal roofing of the skulls is very thin. These features suggest that the Vale specimens are specifically different from those of the Arroyo; but the criteria are at present insufficiently tested to permit absolute interpretation.

### Order Micramphibia

#### Family Gymnarthridae

Fragments of skulls and vertebrae interpreted as gymnarthrids have been found at localities *KD* and *KF*. The identification has been based primarily upon teeth and vertebrae, since no adequately preserved skulls have as yet been found. Generic identification is impossible on the basis of known specimens.

### Order Lysorophia

#### Family Lysorophidae

Lysorophids appear to be quite common in the area, but the only remains which can be definitely assigned to the family are series of vertebrae found in clay nodules incorporated in the conglomerate at locality *KD*. No detailed study has been made, but both the general features of the vertebrae and the mode of occurrence strongly suggest the genus *Lysorophus*. There is a wide range of size in known specimens. In the smallest there appear to be as many as ten vertebrae to the centimeter, while, in the largest, vertebrae are several millimeters in length. Most specimens fall within the size range of typical specimens of *Lysorophus* from the Arroyo. Determination of species in this genus is exceedingly difficult, inasmuch as most available material consists of vertebrae. There is no basis for differentiating the Arroyo and the Vale specimens specifically, but there are no positive criteria for assignment to the same species.

## Class Reptilia

### Subclass Eureptilia

### Infraclass Captorhina

### Order Captorhinomorpha

#### Family Captorhinidae

*Captorhinus* and one specimen which may be tentatively assigned to *Labidosaurus* represent the family Captorhinidae. The best specimen of *Captorhinus* consists of a partial skull and lower jaws, a series of vertebrae, and parts of limb bones. The other specimens are fragmentary. There appears to be no question of generic assignment. The variety of specimens of *Captorhinus* from the Clyde and Arroyo formations of Texas and from Oklahoma has resulted in considerable confusion in the taxonomy of the genus. The Vale specimens fall within the limits of variation witnessed in Arroyo specimens, but it is not possible to make specific assignment until the taxonomy of specimens from the Arroyo has been clarified. The specimen from local-

ity *KD* of the Vale has a median length of about 80 mm., which compares favorably with that of larger specimens from the Arroyo.

The one specimen tentatively referred to *Labidosaurus* consists of a well-preserved snout, including the anterior margin of the orbits. The snout is about 3 inches long and is much like that of *Labidosaurus*, except that the external nares are very large. This character may be sufficient basis for assignment to a new species, but no designation will be made until the captorhinids are thoroughly studied.

### Infraclass Synapsida

### Order Pelycosauria

#### Family Sphenacodontidae

*Dimetrodon* is the only sphenacodont which has been definitely identified from the Vale. This genus is second only to *Diplocaulus* in abundance and has been found as fragments in all localities except *KH* and in quarries in *KA* and *KD*. There is a wide size range in the fragmentary specimens, but that this has taxonomic significance is open to question. All well-preserved specimens seem to belong to a single species. The mature animals were quite large, comparable in size to *D. gigashomogenes* of the Arroyo. No more than fragments of skulls have been found, so that determination must be based upon postcranial elements. The most striking feature of the vertebrae is the exceedingly long neural spines, which are so deeply grooved anteriorly and posteriorly for part of their length that the shafts are nearly separated into lateral halves. Preliminary examination suggests that the Vale *Dimetrodon* is close to *D. gigashomogenes* of the Arroyo. There are differences in the vertebrae which appear to be specifically significant, but it seems probable that the Vale species was derived from *D. gigashomogenes*.

Some of the smaller specimens referred to *Dimetrodon* may well belong to one or more other species, and some fragments may represent other genera. Many of the fragments are so incomplete that identification is difficult.

### Suborder Edaphosauria

#### Family Edaphosauridae

The edaphosaurids of the Vale should offer an interesting series for study when more complete materials have been obtained; but at present they are recognized only from tooth-bearing palatal plates and lower jaws. No trace of the characteristic *Edaphosaurus* neural spines has been found, and this in itself is interesting, since spines are normally abundant in areas where the genus occurs. In view of this, although some of the palatal plates have the characteristic *Edaphosaurus* dental arrangement, there is some doubt that they actually belong to this genus. In addition to pterygoid plates studded with *Edaphosaurus*-like teeth, two large palates with teeth arranged in regular rows have been found. They bear some resemblance to the pterygoid plate from the Arroyo referred tentatively to *Trichasaurus* by A. S. Romer and L. I. Price (1940, p. 423); but generic identity seems improbable, and Romer and Price's reference of the Arroyo plate to *Trichasaurus* is, of course, extremely problematical. There seems little doubt that the large plates from the Vale are edaphosaurid, but more definite assignment is not possible at the present time.

Several partial lower jaws seem to pertain to the edaphosaurid. They are varied in size and may represent more than one genus. None has been found associated with more than a fragment of skull, so

that it has not been possible to determine proper associations.

### INDETERMINATE SPECIMENS

In addition to the specimens which have been discussed, a number have not as yet been identified. This has resulted either from exposure in the field inadequate to permit even rough identification or from an inability to interpret affinities on the basis of structures which are visible. Many of the specimens in the former category may prove to be additional representatives of genera already discussed, but those in the latter group may prove to be of considerable interest as new elements from the Permian. Two of particular interest are discussed in the following paragraphs.

Among the most interesting specimens found is a large scapulo-coracoid whose total height is nearly 15 inches. The coracoid area has been badly damaged by crystallization of gypsum within the bone, and preservation of no part of the bone is adequate. Most puzzling is the extremely narrow scapular blade, which has a maximum width of less than 3 inches. The bone is unlike anything known from the North American Permian. It is too large to be associated with any of the genera from the area, except possibly the large edaphosaurid palatal plate. The resemblance to an edaphosaurid scapulo-coracoid is remote at best.

Parts of several small jaws, 3–4 inches long, have been found. The teeth are conical and sharply pointed, suggesting affinities with some primitive sphenacodont pelycosaur. Such an assignment, however, can be considered only as extremely tentative. In addition to the jaws, there are fragments of limb bones and vertebrae, which give further evidence of the occurrence of small, indeterminate pelycosaurs.

### SUMMARY OF THE FAUNA

The fauna is interesting in a number of respects; and presumably, as details are revealed, they will add a significant chapter to our knowledge of the Permian vertebrates of North America. Positive knowledge shows a fauna similar to that from the Arroyo but containing a few new genera. Possibly because all materials have come from channel-fills, certain genera which are common in the Arroyo are missing. Nonetheless, the absence of certain genera is striking and may prove of evolutionary and stratigraphic importance. Notable is the apparent absence of rhachitomous amphibians. In the lower formation they are not uncommon in stream-channel deposits, and it seems rather remarkable that such large genera as *Eryops* and *Trimerorhachis* have not been encountered. Another notable absence is that of the large parareptilian genus, *Diadectes*, which is abundant in the Arroyo. It is quite possible that future work will reveal the presence of these and other common Arroyo genera, but present indications are that they did not form a part of the fauna.

No definitely new elements have been added to the lower Permian fauna, unless it be the animal represented by the large, undetermined scapulo-coracoid. In almost every respect the fauna has continuity with that of the Arroyo formation and may reasonably be interpreted as an evolutionary development from this fauna.

### NOTES ON THE FLORA

To date, the flora of the deposits has been neglected in large part, since its study was not the primary objective of the work. There is, however, a rather extensive flora present, and it offers considerable promise for future work. Plant re-

mains occur in the channel deposits in direct association with the vertebrates, in green phases of the channels which contain no vertebrates, and in the evenly bedded red and green beds cut by the channels. Many of the plants are preserved as impression only, but in some instances replacement has preserved internal structure. The most common replacing material is hematite. About a dozen plant specimens have been collected, but as yet no study of them has been made.

Most common is a *Calamites*-like plant represented by small to moderately sized stems. Fronds of Filicales or Cycadofilicales are moderately abundant but rarely well preserved. There is some evidence of conifers in strobili and fronds. Evaluation of the flora as a basis for environmental and stratigraphic interpretation must await more extensive collecting and study by an expert in late Paleozoic paleobotany.

ACKNOWLEDGMENTS.—Much credit goes to my field partners in this work, Michael S. Chappars, Roy Reinhart, and Nicholas Hotton III, whose aid in prospecting and collecting made it possible to obtain a good fauna from relatively barren beds in a short time. I also express my appreciation to Mr. Robert Anderson, of the Waggoner Estate, with headquarters at Vernon, Texas, and to Mr. Wallace Nichols, of Dallas, Texas, for granting my parties permission to work upon their respective properties and for extending courtesies which made the work more pleasant and efficient.

## REFERENCE CITED

ROMER, A. S., and PRICE, L. I. (1940) Review of the Pelycosauria: Geol. Soc. America Special Paper 28.

# A NOTE ON THE ORIGINAL ISOTOPIC COMPOSITION OF TERRESTRIAL CARBON[1]

KALERVO RANKAMA
University of Helsinki[2]

ABSTRACT

The primordial isotopic composition of terrestrial carbon is discussed with reference to astrophysical and geological evidence on the possibilities of fractionation of isotopes. It is suggested that the original isotopic composition of terrestrial carbon corresponds to that found in meteorites, where the $C^{12}/C^{13}$ ratio ranges from 89.8 to 92.0. As is known by earlier investigations, the inorganic processes on the earth tend to concentrate the heavy isotope, $C^{13}$, whereas the light one, $C^{12}$, is accumulated by organic processes. The isotopic composition of O, Fe, and Cu, as compared with that of C, in meteorites and terrestrial samples is discussed.

## INTRODUCTION

There is a very definite trend in present-day geology and related sciences toward considering the earth as a vast physicochemical system, treating the facts collected statistically and using the results obtained from the study of a number of more or less limited areas on a global scale. Geology, as a science, has decidedly entered a period of "exactization" and today applies the results and methods of the more fundamental sciences: physics, chemistry, and astronomy. Among other things the more recent results in the study of isotopes are now being applied to geology and geochemistry. Wickman (1941) was the first to use isotopes in the study of a geochemical problem. He calculated the total amount of coal and bitumen on the basis of the abundance ratios of the carbon isotopes $C^{12}$ and $C^{13}$, determined by Nier and Gulbransen (1939) and by Murphey and Nier (1941). Another contribution has been made by the present author in a still unpublished study of the origin of pre-Cambrian carbon. For that investigation all the original $C^{12}/C^{13}$ determinations carried out by Murphey (1941) were made available to the author through the courtesy of Dr. Alfred O. C. Nier. They proved to be of very considerable geochemical interest, and it was found possible to use them as a starting-point for a discussion dealing with the original isotopic composition of terrestrial carbon, which will be presented in this paper. However, it is necessary to consider recent results of astrophysics and geology, which serve as a useful basis for the study of the isotope fractionation.

[1] Manuscript received January 23, 1948.

[2] On leave; at present at the University of Chicago.

## ASTROPHYSICAL AND GEOLOGICAL BACKGROUND OF THE FRACTIONATION OF CARBON ISOTOPES

### GENERAL REMARKS

Although several determinations of the isotope ratio $C^{12}/C^{13}$ were made by previous workers, the investigations of Nier, Gulbransen, and Murphey referred to above are the first systematic studies of the variation of the $C^{12}/C^{13}$ ratio in different sources of carbon found in nature. They found that the heavy isotope, $C^{13}$, is concentrated in limestones, whereas carbon of vegetable origin has a preference for the light isotope, $C^{12}$. The variations in the $C^{12}/C^{13}$ ratio are thus related to the origin of the samples, and Nier and

his associates assumed that the igneous carbons represent the original $C^{12}/C^{13}$ ratio. The opposite conclusion was reached by Wickman (1941, p. 420), who pointed out that, in view of the modern petrogenetic theories, carbon found in igneous rocks could not be considered to prove anything about the original $C^{12}/C^{13}$ ratio. Similarly, he concluded that the meteorites could not supply an answer to this question because of their variation of the $C^{12}/C^{13}$ ratio, which is greater than that found in any other group analyzed by Nier and his co-workers. Wickman finally assumed that diamond showed the original isotopic composition of terrestrial carbon and pointed out that the corrected $C^{12}/C^{13}$ value of this mineral was lower than any found in the meteorite carbons—a fact not at all unreasonable, considering the wide variation in the latter case.

### ASTROPHYSICAL BACKGROUND

The very fragmentary information which is collected on the interior structure of the earth is based mainly on the propagation of seismic waves and on meteorite studies. It is commonly believed today that the iron meteorites correspond to the hypothetical nickel-iron core of the earth, whereas the silicate meteorites are thought to be counterparts of the upper layers of the earth's crust, notably the Sima layer. The origin of the meteorites has been a matter of controversy for some time, but, as stated by Paneth (1940, 1946), the evidence is now believed to be overwhelmingly in favor of meteorites' being fragments of a member of the solar system. Similarly, petrological properties of the meteorites point strongly toward their origin by fragmentation of a larger astronomical body, as is stated by Foshag (1941, p. 144). The view that the meteorites are scattered parts of a single celestial body was first expressed more than a hundred years ago by the German astronomer, Heinrich Wilhelm Matthias Olbers. Much trouble was caused by the fact that the most recent age determinations of siderites, carried out by Paneth and his co-workers by the use of the helium method, gave values ranging from less than $0.11 \cdot 10^6$ years to $6{,}800 \cdot 10^6$ years (Arrol, Jacobi, and Paneth, 1942). On the basis of these values they claimed that the solar system cannot be less than $7{,}000 \cdot 10^6$ years old. On the other hand, there is much astronomical and astrophysical evidence indicating that the age of the universe is of the order of $3{,}000 \cdot 10^6$ years, which value was given by Chandrasekhar (1944) and by Shapley (1945). There are, accordingly, two time scales for the universe, and most of the evidence now available seems to be in favor of the shorter. The varying helium ages of the iron meteorites are explained by Bauer (1947) in a recent contribution in which he emphasizes the fact that the high helium contents appear only in the relatively small masses. The explanation offered by Bauer is that cosmic radiation has produced extra helium in the small meteoroids. Nuclear disruptions in which $\alpha$-particles are among the disintegration products are caused by cosmic radiation. Bauer has also calculated the rate of $\alpha$-particle formation for a meteoroid, and he concludes that the maximum helium content observed is produced in $3{,}400 \cdot 10^6$ years, which is just half the highest age calculated by Paneth.

It may thus be concluded that the most recent results of the age determination of iron meteorites furnish further proof of the validity of the short time scale. Shapley's (1945, p. 520) opinion is that the evidence of the moment is fairly strong that the planets, the meteorites,

the galaxies, the double stars, possibly the clusters of galaxies, and certainly the expansion of the galaxies in the part of the metagalaxy which can be observed all date back to a time about $3{,}000 \cdot 10^6$ years ago when something momentous happened, or to the zero-hour epoch, $t_0$, in the expansion of the universe of galaxies.

### GEOLOGICAL BACKGROUND

According to Shapley (1945, p. 511), the astronomers see little support for the suggestion that all the planets of our solar system may not have been born together. Shapley also thinks (1945, p. 508) that the necessarily rapid cooling and crustal solidification of small or medium-sized astronomical bodies caused the congealing of the earth's crust at a time when the galaxies were still closely packed together. According to the calculations presented in the previous section, the age of the earth should thus be approximately $3{,}000 \cdot 10^6$ years or, to use the value determined by Bauer for the iron meteorites, $3{,}400 \cdot 10^6$ years.

A new value for the age of the earth, of about $3{,}350 \cdot 10^6$ years, was recently presented by Holmes (1947). This is a revised estimate based on Nier's analyses of the isotopic composition of lead from galenas and other lead minerals of known geological age. The close correspondence between the above value and that given by Bauer for the length of helium production in siderites is, perhaps, more than a strange coincidence. According to Ahrens (1948), the age of the pegmatites of southeastern Manitoba, as determined by the use of the strontium method, is about $2{,}100 \cdot 10^6$ years. Although these pegmatites are evidently the oldest rocks whose age has ever been measured, the surrounding rock complex, by its geological properties, must be still more ancient, although it certainly does not represent the primordial crust of the earth. Certainly, the first crust of our planet is nowhere exposed, and even the oldest rocks thus far encountered must be several hundred, if not a thousand, million years younger than the first patches of solid crust which quite probably were remelted several times after their formation, until a crust of sufficient thickness and permanence surrounded the earth. It may also be possible, as suggested by the present writer (Rankama, 1946), that the upper lithosphere of today is the product of a continuous process of differentiation, whereby the elements peculiar to granitic rocks are carried and concentrated upward, while other elements are pushed back toward the basaltic substratum. Therefore, it is impossible to make any direct determination of the absolute astronomical age of our planet based on the use of geological materials. On the other hand, it is safe to assume that if age determinations were made of the deeper parts of the earth, they would give higher values than those obtained for the superficial parts of the earth. Attention must also be paid to the possibility that changes in the isotopic composition of elements, such as would be likely to affect the results of the age determinations, might take place during the processes outlined above.

It may be concluded from the discussion presented above that present-day evidence favors the coevality and common origin of the earth and the meteorites. Therefore, the discussion on the fractionation of the carbon isotopes which follows is based on the principle of the similarity of meteoritic and terrestrial matter. The uppermost parts of the lithosphere are not as yet known to have a counterpart among the meteorites.

### COSMIC ABUNDANCE OF $C^{12}$ AND $C^{13}$

During the sidereal and planetary life of the elements their isotopic composition is evidently subject to changes, and the present abundance ratios of the isotopes are the result of a long evolution. The primeval carbon, in particular, is thought to possess a $C^{12}/C^{13}$ ratio different from that found in present-day carbonaceous substances.

The presence of the carbon isotopes in class N stars was established by Sanford (1929). Menzel (1930, p. 35) concluded that $C^{12}$ is certainly not more than ten times as abundant as $C^{13}$ in the N-type stars. Fairly convincing evidence of the presence of $C^{13}$ in R stars also was found by Sanford (1932).

The occurrence of the carbon isotopes $C^{12}$ and $C^{13}$ in the spectra of N-type stars was further investigated by Shajn (1942), who found the relative concentration of $C^{13}$ to be between 0.05 and 0.50, depending on the particular N star, or much in excess of the values recorded on the earth. The corresponding $C^{12}/C^{13}$ ratios, calculated from the values given by Shajn, are 19 and 1, respectively. Recently, McKellar (1947) presented tentative values for the ratio $C^{12}/C^{13}$ in fifteen R-type stars. McKellar was able to distinguish at least two groups among the R stars, one with $C^{12}/C^{13}$ equal to 50 or more and another group with a ratio of about 3.4. He also suggested that in the first group the composition of the carbon was, perhaps, not unlike terrestrial and meteoritic samples where the ratio is about 90. It should be pointed out that very strong bands due to carbon molecules and carbon compounds are present in the spectra of the R and N stars. Hence it is assumed that the atmospheres of these stars are reducing because the amount of carbon present therein is higher than that of oxygen. Most of the cooler stars, on the other hand, have oxidizing atmospheres, in which the excess of oxygen permits the formation of oxides.

A summary of the evidence available on the relative abundance of the carbon isotopes $C^{12}$ and $C^{13}$ in stellar atmospheres shows beyond all doubt that there are very pronounced differences in the $C^{12}/C^{13}$ ratio of the N- and the R-type stars. As pointed out by McKellar (1947), this distinction is of interest in relation to stellar evolution and nuclear reactions taking place in the stars. According to Bethe (1939), $C^{12}$ and $C^{13}$ play an important part as catalysts in the nuclear reactions producing helium from hydrogen. Therefore, the stage of the more stable $C^{12}/C^{13}$ ratio cannot be initiated until a phase in the stellar evolution is reached where the possibility of nuclear reactions is excluded.

Furthermore, Bobrovnikoff (1930) has established the presence of $C^{13}$ in comets; and, according to Swings (1943, p. 91), the relative abundance of $C^{12}$ and $C^{13}$ may be different in various comets.

### TERRESTRIAL AND METEORITIC ABUNDANCE OF $C^{12}$ AND $C^{13}$

King (1936) and Jenkins and King (1936) were the first to determine the abundance of $C^{13}$ in a meteorite. Their results, obtained by the use of spectrographic methods, showed no difference in the relative abundance of the isotopes $C^{12}$ and $C^{13}$. Murphey (1941) analyzed several meteorites, the average $C^{12}/C^{13}$ ratios being published by Murphey and Nier (1941). As to terrestrial samples, the reader is referred to the papers mentioned on page 199. There is, in addition, a paper by West (1945) which deals with the relative abundance of $C^{12}$ and $C^{13}$ in petroleum. The total number of individual $C^{12}/C^{13}$ determinations now available

is, therefore, more than one hundred. The values, a number of which are given as averages, are listed in table 1.

The values of table 1 show that there are systematic differences among the four existing sets of data. The ratios given by Murphey are higher than the previous ones published by Nier and Gulbransen, as was pointed out by Wickman (1941) and by West (1945), whereas the values of the latter are still higher and those given by the present author are comparable to Murphey's figures. As three different mass spectrometers have been used to obtain the values listed in table 1, it is considered appropriate to exclude, in the following, the values given by Nier and Gulbransen and by West, for reasons discussed in a previous paper (Rankama, in press). The remaining values, which were obtained with the same mass spectrometer, are presented in figure 1, the $C^{12}/C^{13}$ ratios being therein grouped according to the origin of the samples investigated.

Starting from the fact established by Nier and Gulbransen (1939) that plants tend to concentrate the lighter isotope, $C^{12}$, whereas the heavier one, $C^{13}$, becomes concentrated in limestones, the presence of two groups of carbon, inorganic and organic, can be established.[3] However, neither of these processes seems to be able to lead to a complete separation of the carbon isotopes during the fractionation. In some cases the $C^{12}/C^{13}$ ratios would also seem to be influenced by the presence of both inorganic and organic carbon, as in the case of metamorphosed sediments which may contain graphite deposited initially as a residual mineral, together with carbon of bituminous origin. Likewise, as stated by Murphey (1941), the carbonates deposited from sea water containing decaying humic matter would consist of carbon of lower $C^{13}$ percentage than those precipitated in water free from such contamination. An example of such conditions is actually furnished by the $C^{12}/C^{13}$ ratio in the carbonate from the coal ball (table 1).

[3] Additional evidence that the content of $C^{13}$ in animal tissues is lower than in mineral sources was furnished by Swendseid, Barnes, Hemingway, and Nier (1942).

### THE PRIMORDIAL ISOTOPIC COMPOSITION OF TERRESTRIAL CARBON

The position occupied by the meteoritic carbons in figure 1 is of considerable interest. The $C^{12}/C^{13}$ ratio of meteoritic carbon is 89.8–92.0, but that of inorganic carbon is 87.9–90.2, and of the organic group it is 90.3–93.1. The meteoritic carbon evidently forms an intermediate group between the carbons of inorganic and those of organic origin. Therefore, it seems reasonable to assume that the meteoritic carbon does actually show the primordial isotopic composition of this element, in so far as our solar system is considered, that is, at a stage when the cooling excluded further nuclear reactions. Such a conclusion would fit very well into the general picture of the fractionation of the carbon isotopes because, starting from the original $C^{12}$ and $C^{13}$ contents, there are two opposite series of processes, viz., the inorganic processes tending to concentrate $C^{13}$ and lower the $C^{12}/C^{13}$ ratio and the organic ones accumulating $C^{12}$ and increasing the value of the isotope ratio. Similarly, the astrophysical and geological background of the fractionation presented above gives no evidence which would contradict this assumption. It should be noted, in addition, that during the birth of the earth no geological or organic processes were active which would have been able to modify the primordial isotopic composition of carbon in the upper parts of the

## TABLE 1

### $C^{12}/C^{13}$ Ratios in Carbonaceous Materials

| Sample and Location | $C^{12}/C^{13}$ | Reference* |
|---|---|---|
| **Carbon from Meteorites** | | |
| Homestead, Iowa (*Cgb*) | 91.7 | M |
| Estherville, Iowa (*M*) | 91.1 | M |
| Finnmarken, Norway (*P*) | 91.3 | M |
| Pavlodar, Semipalatinsk, Siberia, U.S.S.R. (*P*) | 92.0 | M |
| Cañon Diablo, Ariz. (*Og*) | 89.8 | M |
| Cosby's Creek, Tenn. (*Og*) | 89.4 | NG |
| Cosby's Creek, Tenn. (*Og*) | 91.6 | M |
| Kokomo, Ind. (*Iron*) | 91.3 | M |
| **Igneous Carbon** | | |
| Diamond: Kimberley, Union of South Africa | 89.0 | NG |
| Graphite: Ceylon | 89.8 | NG |
| Graphite: Ceylon | 90.2 | M |
| Calcite (with zeolite): Patterson, N.J. | 89.9 | M |
| Calcite: Joplin, Mo. | 89.4 | M |
| Calcite from nepheline pegmatite: Bancroft, Ont., Canada | 89.8 | M |
| **Calcium Carbonate and Limestones** | | |
| Clam shell: Boston, Mass. (Recent) | 88.7 | NG |
| Marine shell | 89.5 | M |
| Limestone: Bermuda | 89.0 | M |
| Limestone: Teplitz, Bohemia (upper Cretaceous) | 89.3 | M |
| Chalk | 89.2 | M |
| Limestone: Bavaria (Jurassic) | 89.2 | M |
| Limestone (Mississippian) | 89.0 | M |
| Limestone: New Salem, N.Y. (lower Devonian) | 89.2 | M |
| Limestone: Champlain Valley, N.Y. (mid-Ordovician) | 88.8 | M |
| Limestone: Vermont (Ordovician) | 88.6 | NG |
| Limestone: northwestern Vermont (upper Cambrian) | 89.4 | M |
| Limestone: Frontenac Co., Canada (pre-Cambrian) | 89.3 | M |
| Limestone: Champlain Valley, N.Y. (pre-Cambrian) | 89.3 | M |
| Limestone: New York (Grenville) | 89.3 | M |
| Limestone: New York (Grenville) | 87.9 | NG |
| Graphite from limestone: Parainen, Finland (Archean) | 88.4 | R |
| **Air and Water** | | |
| Air: Massachusetts, 3/14/38 | 92.5 | NG |
| Air: Massachusetts, 3/22/38 | 89.9 | NG |
| Air: Minneapolis, Minn., 3/1/41 | 91.5 | M |
| Carbon dioxide from soil: Tulsa, Okla. | 90.5 | W |
| Carbon dioxide from soil: Tulsa, Okla. | 90.1 | W |
| Sea water | 89.3 | M |
| **Bituminous Sediments** | | |
| Bottom ooze: Mud Lake, Fla. (Recent) | 91.6 | M |
| Oil shale: Elko, Nev. (Miocene or Pliocene) | 92.6 | M |
| Kuckersite: Esthonia (Ordovician) | 92.7 | M |
| Kuckersite No. 2: Esthonia (Ordovician) | 92.5 | M |
| Kolm: Sweden (upper Cambrian) | 92.4 | M |
| Shungite: Shunga, K.F.S.S.R. (pre-Cambrian) | 92.7 | M |
| Shungite, 1st variety: Shunga, K.F.S.S.R. (pre-Cambrian) | 92.9 | R |
| Graphite from carbon-bearing schist: Paakki, Paltamo, Finland (Archean) | 90.3 | R |
| Graphite: Kärpälä, Mäntyharju, Finland (Archean) | 91.7 | R |
| Graphite from graphite phyllite: Kalkkimaa, Kemi, Finland (Archean) | 90.5 | R |
| Carbon from slate fragment: Valkeekivi, Ylöjärvi, Finland (Archean) | 91.0 | R |
| Carbon from *Corycium enigmaticum:* Tähtinen, Aitolahti, Finland (Archean) | 90.8 | R |
| Carbon from *Corycium enigmaticum:* Tähtinen, Aitolahti, Finland (Archean) | 92.0 | R |
| Carbon from *Corycium*-like inclusion: Saarinen, Aitolahti, Finland | 91.4 | R |
| Albertite: Nova Scotia, Canada | 92.4 | M |
| Green River oil shale: Rulison, Colo. | 92.5 | M |
| Kerosene shale: Australia | 91.7 | M |
| Alum shale (Alaunerde): Germany | 92.2 | M |
| **Petroleum and Natural Gas** | | |
| Crude oil: Allen, Okla. | 92.8 | M |
| Crude oil: Rusk Co., Tex. | 92.3 | M |
| Crude oil: Los Angeles, Calif. | 92.0 | M |
| Oil: Emba, U.S.S.R. | 92.5 | M |
| Oil: Western Texas | 91.2 | NG |
| Oil and gas: Barton and Rice Counties, Kan. | 94.1 | W |
| Oil: Pottawatomie Co., Okla. | 93.2 | W |
| Oil: Winkler Co., Tex. | 93.9 | W |
| Oil: Carbon Co., Wyo. | 94.1 | W |
| **Carbon of Animal Origin** | | |
| Clam flesh: Boston, Mass. | 90.1 | NG |
| Cod oil | 92.1 | M |
| Casein | 92.5 | M |
| Gelatin | 91.4 | M |
| **Carbon of Vegetable Origin** | | |
| Linseed oil | 92.8 | M |
| Raw China wood oil | 92.8 | M |
| Rubber | 92.5 | M |

* References: "NG," Nier and Gulbransen (1939); "M," Murphey (1941); "W," West (1945); "R," Rankama (in press).

TABLE 1—*Continued*

| SAMPLE AND LOCATION | $C^{12}/C^{13}$ | REFERENCE* |
|---|---|---|
| | Carbon of Vegetable Origin | |
| Spores of *Lycopodium clavatum* | 93.1 | M |
| "Balkhashite" algae: Turkestan, U.S.S.R. | 92.8 | M |
| Weed: Tulsa, Okla. | 92.6 | W |
| Pine wood: Massachusetts | 91.5 | NG |
| Pine wood | 91.6 | M |
| Oak wood | 91.6 | M |
| Maple wood: Massachusetts | 91.9 | M |
| Peat: Frostproof, Fla. | 92.8 | M |
| Interglacial wood: Kittson Co., Minn. | 92.2 | M |
| Wood: Moorhead, Minn. (late Pleistocene) | 91.7 | M |
| Wood (Pleistocene) | 91.9 | M |
| Spruce wood: Springfield, Minn. (Pleistocene) | 91.8 | M |
| Lignite: Siebenberg, Germany (Pliocene-Miocene) | 91.8 | M |
| Xyloid lignite: Brandon, Vt. (Eocene) | 91.7 | M |
| Underclay of coal: Golden, Colo. (Cretaceous) | 91.9 | M |
| Lignite: Pastry, France (Cretaceous) | 91.9 | M |
| Bituminous coal: Carillos, N.M. (Cretaceous) | 91.5 | M |

| SAMPLE AND LOCATION | $C^{12}/C^{13}$ | REFERENCE* |
|---|---|---|
| | Carbon of Vegetable Origin | |
| Jet: Whilby, England (Jurassic) | 91.3 | M |
| Spore coal: East Liverpool, Ohio (Carboniferous) | 92.6 | M |
| Cannel coal: Linton, Ohio (Carboniferous) | 92.1 | M |
| Bituminous coal: Coal City, Ill. (Carboniferous) | 92.0 | M |
| Bituminous coal: Waukee, Iowa (Carboniferous) | 91.6 | M |
| Torbanite, black: Scotland (Carboniferous) | 92.0 | M |
| Torbanite, brown: Scotland (Carboniferous) | 91.3 | M |
| Semianthracite: Antarctica | 92.3 | M |
| Anthracite: Kingston, Pa. (Carboniferous) | 91.4 | M |
| Anthracite: Tennessee (Carboniferous) | 91.8 | NG |
| Graphitized anthracite: Mansfield, Mass. | 91.8 | M |
| Coal ball, organic: Dallas Co., Iowa | 91.8 | M |
| Coal ball, inorganic (carbonate): Dallas Co., Iowa | 90.6 | M |

lithosphere. Therefore, it seems to be safe to assume that the original isotopic composition of carbon cannot be that found in the present-day carbons of inorganic or organic origin.

The reaction mechanisms responsible for the fractionation of the carbon isotopes are, as yet, known only incompletely. Isotopic exchange reactions certainly play an important part in the fractionation. Urey and Greiff (1935) have suggested the possibility of at least some separation's taking place by means of the reaction between carbon dioxide and the bicarbonate ion. This exchange reaction is evidently responsible for the concentration of the heavier isotope in carbonate sediments. The separation mechanisms active in connection with igneous phenomena might include diffusion in the gravitational field and in a pressure field, distillation, evaporation, and chemical separations. As to the mechanisms causing the concentration of the light isotope, $C^{12}$, in animals and plants, it seems to be evident that the organisms form special cases of the fractionation caused by physicochemical systems. Kamen (1946, p. 121) admits that isotope differentiation may be more characteristic of living systems than of inert systems. With special reference to the enrichment of $C^{12}$ in plants, it should be noted that recent investigations of Armstrong and Schubert (1947) indicate that the exchange between carbon dioxide and insoluble carbonates in the leaves is a probable method by which green plants absorb carbon dioxide from the air. These authors have found that a definite exchange of $C^{13}$ occurs between atmospheric $CO_2$ (containing about 1.1 per cent $C^{13}$ [Belkengren, Nier, and Burr, 1942, Nature, vol. 149, p. 24]) and enriched $C^{13}$ barium carbonate.

The foregoing observations furnish, in addition, some information concerning the cycle of the carbon isotopes between the atmosphere and the sea. This cycle is diagrammatically presented in figure 2, and the reader is referred to the $C^{12}/C^{13}$ values given in table 1. As was pointed out above, $C^{13}$ is enriched in sea water during the exchange reaction between atmospheric carbon dioxide and the bicarbonate ion. During the formation of the carbonate sediments by inorganic or organic precipitation there probably oc-

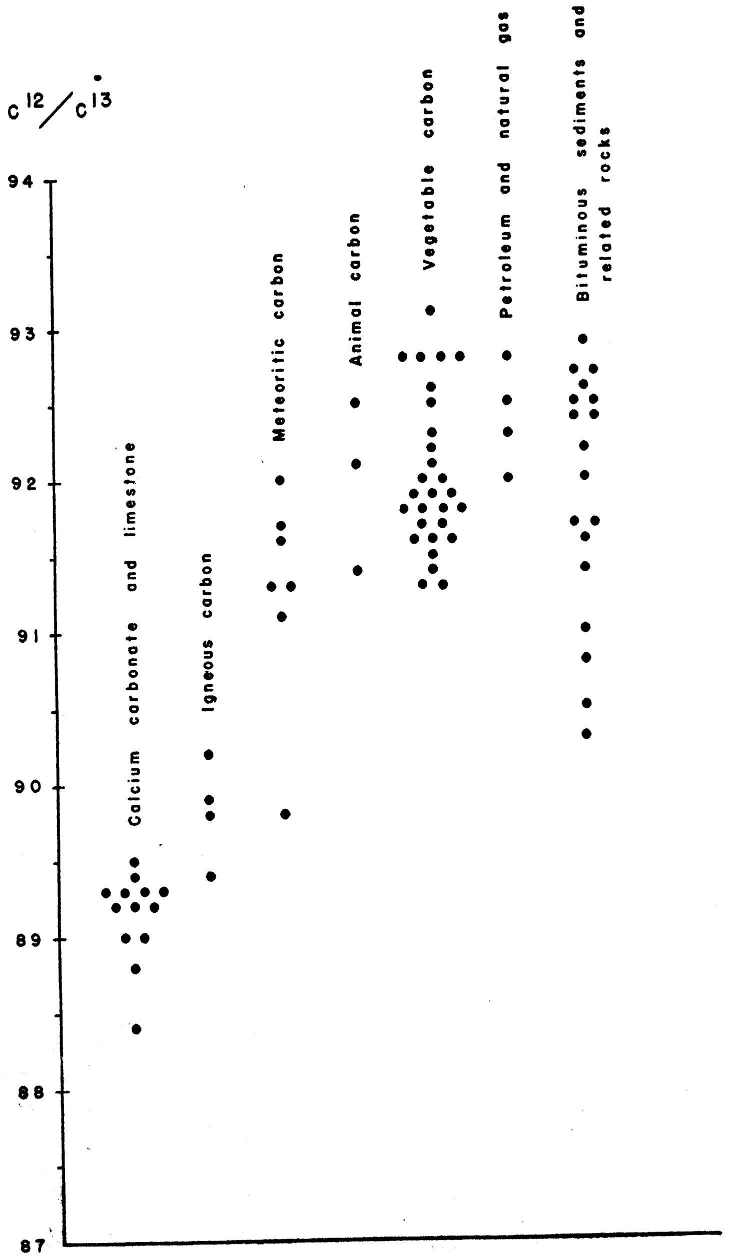



FIG. 1.—$C^{12}/C^{13}$ ratios in carbonaceous materials, classified according to their origin

curs further enrichment of $C^{13}$, whereas the content of this isotope in the carbonate sediments is bound to decrease when the sediments are brought into contact with the atmospheric carbon dioxide. The light isotope, $C^{12}$, on the other hand, becomes enriched in marine plants and, consequently, in marine animals feeding upon the plants. During the decay of the organisms a part of the carbon may be carried to the atmosphere as carbon dioxide. As yet, there is no evidence available to show whether or not $C^{12}$ will become further enriched during this process.

As to the animals, the reaction mechanisms seem to be known very inadequately. Belkengren (according to Murphey, 1941) has found that fats reject $C^{13}$ more actively than do any other substances encountered.

Still another fact seems to be of certain interest in this connection. In general, the inorganic processes causing the fractionation of the carbon isotopes seem to be slow as compared with the more rapid organic processes taking place in nature. In any case, these two processes work continuously in opposite directions; and, to judge by figure 1, the inorganic processes have thus far caused a more thoroughgoing fractionation than have the organic ones. The petroleum hydrocarbons might form a possible exception if the values presented by West are considered, but here the effects of distillation and evaporation may already be effective.

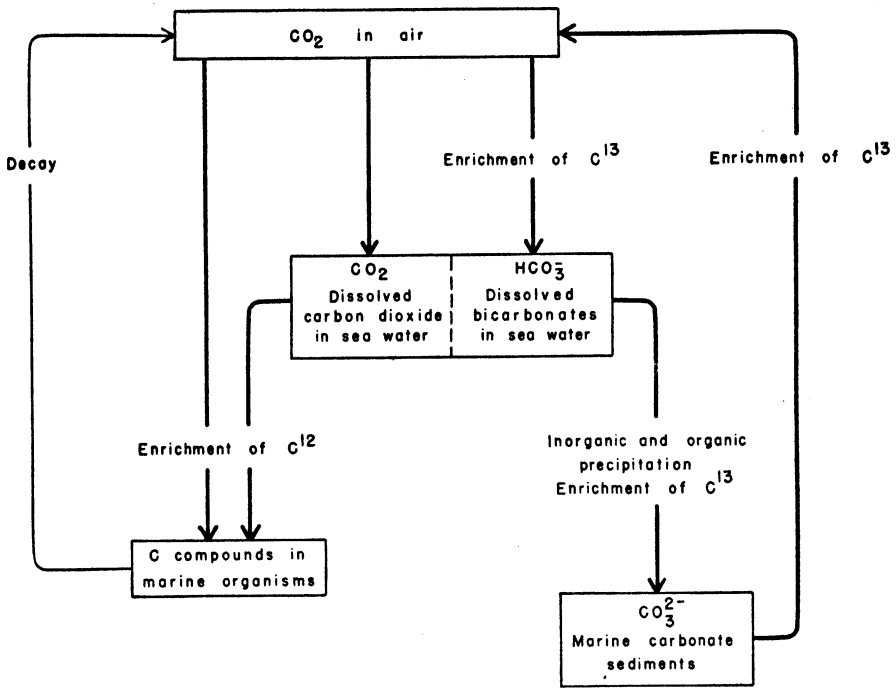



FIG. 2.—Cycle of $C^{12}$ and $C^{13}$ between the atmosphere and the sea

### SOME OTHER ABUNDANCE STUDIES OF ISOTOPES BASED ON METEORITES

The oxygen, iron, and copper isotopes have been studied more recently in meteorites, in order to establish the presence of possible variations in their abundance ratios. Manian, Urey, and Bleakney (1934) reported no variations in the isotope ratio $O^{16}/O^{18}$ of stony meteorites as compared with terrestrial oxygen, although they considered that a higher concentration of $O^{17}$ in meteorites might be possible (1934, p. 2608). Valley and Anderson (1941) studied the iron isotopes $Fe^{54}$, $Fe^{57}$, and $Fe^{58}$. Even though the averages published by these authors consistently show slightly lower abundance values of the above isotopes in iron meteorites, as compared with irons of terrestrial origin, it is concluded that there is no significant difference in the composition of the samples. The material used by Valley and Anderson consisted of seven specimens of terrestrial iron ores and twelve meteorites. Unfortunately, these authors did not publish any details concerning the source of their material or the results of the separate determinations. Still more recently, Brown and Inghram (1947) studied the isotopic composition of meteoritic (one sample) and terrestrial copper (two samples). The slight difference found to exist in the $Cu^{63}/Cu^{65}$ ratio is thought to be within the experimental error. However, these authors have based their conclusions on a very limited material, and evidently more determinations are needed to prove the validity of their assumption. By reference to table 1 and figure 1 it may be claimed, by using just one of the meteoritic $C^{12}/C^{13}$ ratios, that the isotopic composition of meteoritic carbon does not materially differ from the composition of the carbon of petroleum or of calcite or that there is a very pronounced difference between meteoritic carbon and the carbon of limestones. However, even in the case of carbon, the seventy-odd samples used to construct figure 1 still cannot furnish all the details of the fractionation.

Another point of view is that, if there are variations in the isotopic composition of Fe, Cu, and other elements, they may be too small to be detected by the present-day mass spectrometric techniques. Here also the general geochemical character of these elements may enter as a deciding factor. Iron is very typically siderophile, although it also possesses chalcophile, lithophile, and biophile trends. Copper is mainly chalcophile, but carbon is geochemically very versatile, being siderophile, clearly lithophile, typically atmophile (as $CO_2$), and, above all, a decidedly biophile element. In this case the possibility of a pronounced fractionation might be greater than in the case of many other elements geochemically more constant than carbon.

## REFERENCES CITED

AHRENS, L. H. (1947) Geological age: the extreme antiquity of pegmatites from Manitoba: Nature, vol. 160, p. 874.

ARMSTRONG, W. D., and SCHUBERT, JACK (1947) Exchange of carbon dioxide between barium carbonate and the atmosphere: Science, vol. 106, p. 403.

ARROL, W. J.; JACOBI, R. B.; and PANETH, F. A. (1942) Meteorites and the age of the solar system: Nature, vol. 149, p. 235.

BAUER, CARL AUGUST (1947) Production of helium in meteorites by cosmic radiation: Phys. Rev., vol. 72, p. 354.

BETHE, H. A. (1939) Energy production in stars: Phys. Rev., vol. 55, p. 434.

BOBROVNIKOFF, N. T. (1930) Carbon isotopes in comets: Astron. Soc. Pacific Pub. 42, p. 117.

BROWN, HARRISON, and INGHRAM, M. G. (1947) The isotopic composition of meteoritic copper: Phys. Rev., vol. 72, p. 347.

CHANDRASEKHAR, S. (1944) Galactic evidences for the time-scale of the universe: Science, vol. 99, p. 133.

FOSHAG, W. F. (1941) Problems in the study of meteorites: Am. Mineralogist, vol. 26, p. 137.

HOLMES, ARTHUR (1947) A revised estimate of the age of the earth: Nature, vol. 159, p. 127.

JENKINS, F. A., and KING, ARTHUR S. (1936) A test of the abundance of the heavy isotope carbon in a graphite meteorite: Astron. Soc. Pacific Pub. 48, p. 323.

KAMEN, MARTIN D. (1946) Survey of contemporary knowledge of biogeochemistry. 1. Isotopic phenomena in biogeochemistry: Am. Mus. Nat. History Bull. 87, p. 101.

KING, ARTHUR S. (1936) A spectroscopic examination of meteorites: Astrophys. Jour., vol. 84, p. 507.

MANIAN, SAMUEL H.; UREY, HAROLD C.; and BLEAKNEY, WALKER (1934) An investigation of the relative abundance of the oxygen isotopes $O^{16}:O^{18}$ in stone meteorites: Am. Chem. Soc. Jour., vol. 56, p. 2601.

MCKELLAR, ANDREW (1947) Intensity measurements on the main and isotopic carbon bands in spectra of the R-type stars: Astron. Soc. Pacific Pub. 59, p. 186.

MENZEL, DONALD H. (1930) The identification and cosmic abundance of carbon isotopes: *ibid.*, vol. 42, p. 34.

MURPHEY, BYRON FREEZE (1941) The relative abundances of the isotopes of carbon and oxygen, M.A. thesis, Univ. of Minnesota.

——— and NIER, ALFRED O. (1941) Variations in the relative abundance of the carbon isotopes: Phys. Rev., vol. 59, p. 771.

NIER, ALFRED O., and GULBRANSEN, EARL A. (1939) Variations in the relative abundance of the carbon isotopes: Am. Chem. Soc. Jour., vol. 61, p. 697.

PANETH, F. A. (1940) The origin of meteorites, Halley lecture at Oxford.

——— (1946) Meteorites, and the age of the solar system: Monthly Astron. Newsletter 36, p. 9.

RANKAMA, KALERVO (1946) On the geochemical differentiation in the earth's crust: Comm. géol. Finlande Bull. 137.

——— (1948) New evidence of the origin of pre-Cambrian carbon: Geol. Soc. America Bull. (in press).

SANFORD, R. F. (1929) Carbon isotopes in class N stars: Astron. Soc. Pacific Pub. 41, p. 271.

——— (1932) Further details ascribable to the bands of the carbon isotope $C^{13}$ in stars of spectral classes R and N: *ibid.*, 44, p. 246.

SHAJN, G. A. (1942) The occurrence of carbon isotopes in the spectra of N-type stars: Abastumani Astrophys. Observatory on Mount Kanobili (U.S.S.R.), Bull. 6; (abstr.): Monthly Astron. Newsletter 5, 1942.

SHAPLEY, HARLOW (1945) On the astronomical dating of the earth's crust: Am. Jour. Sci., vol. 243-A, Daly Vol., p. 508.

SWENDSEID, MARIAN E.; BARNES, RICHARD H.; HEMINGWAY, ALLAN; and NIER, A. O. (1942) Formation of acetone bodies from acetic acid: Jour. Biol. Chemistry, vol. 142, p. 47.

SWINGS, P. (1943) Cometary spectra: Royal Astron. Soc. Monthly Notices 103, p. 86.

UREY, HAROLD C., and GREIFF, LOTTI J. (1935) Isotopic exchange equilibria: Am. Chem. Soc. Jour., vol. 57, p. 321.

VALLEY, G. E., and ANDERSON, H. H. (1941) The relative abundance of the stable isotopes of terrestrial and meteoritic iron: Phys. Rev., vol. 59, p. 113.

WEST, S. S. (1945) The relative abundance of the carbon isotopes in petroleum: Geophysics, vol. 10, p. 406.

WICKMAN, FRANS E. (1941) On a new possibility of calculating the total amount of coal and bitumen: Geol. Fören. Stockholm Förh., vol. 63, p. 419.

# GEOLOGICAL SIGNIFICANCE OF SURFACE TENSION[1]

JEAN VERHOOGEN
University of California, Berkeley

ABSTRACT

Surface tension is found to be a controlling factor in a number of processes of geological importance: precipitation and solution in small pores and cavities (metasomatism), incipient crystallization, recrystallization and formation of porphyroblasts, vapor tension of a magma, development of pressure by reheating of rocks containing adsorbed water, and possibly surficial aspects of lava flows.

## INTRODUCTION

It has not been generally recognized that surface phenomena may play an important part in determining the course of geological processes and the state of equilibrium of geological systems. Bain (1936) apparently was the first to make use of the laws of surface tension and adsorption in the study of metasomatism. Besides this pioneer work and with the exception of a few scattered references, the subject has been usually ignored by geologists and is rarely mentioned in textbooks on geology. The present paper points out a few types of geological processes in which surface phenomena presumably play a part. It is offered in the hope that more attention will be given to this type of phenomenon.

## SYMBOLS

The following symbols are used consistently throughout the paper: $T$ is temperature, $P$ pressure, $N$ molar fraction. Lower subscripts refer to a constituent, upper symbols to a phase. Thus $N_i^a$ is the molar fraction of constituent $i$ in phase $\alpha$, and $n_i^a$ is the number of moles of constituent $i$ in this phase, so that $N_i^a = n_i^a/\sum_i n_i^a$ and $\sum_i N_i^a = 1$. The quantity $\mu_i$ is a chemical potential, $p_i^*$ a fugacity, $p_i$ a vapor pressure, and $f_i$ an activity coefficient, the relations between these quantities being

$$p_i^* = p_i f_i = PN_i f_i \,, \qquad \lim_{P \to 0} f_i = 1 \,.$$

The symbols $V^a$ and $S^a$ refer, respectively, to the total volume and total entropy of phase $\alpha$, and $v^a$ and $s^a$ are, respectively, molar volumes and molar entropies. The symbol $\bar{v}_i^a$ is the partial molar volume of $i$ in phase $\alpha$; $a$ is an area, $r$ a radius of curvature, $\sigma$ a surface tension, and $\Gamma_i^a$ the adsorption of $i$ at the interface $a$. The symbol $S^a$ is the total entropy of the surface phase $a$, and $s^a$ its entropy per unit area; $R$ is the gas constant per mole (83.15 bar cm³/1°).

## FUNDAMENTAL RELATIONS

The *surface energy* of an interface between two homogeneous phases is a measure of the increase of energy acquired by a molecule which has come from the interior of one of the phases to form a new unit of surface. The *surface tension* $\sigma$ is usually defined by the relation $dW = \sigma da$, where $dW$ is the work expended on increasing the area of the interface by an amount $da$. Although these two quantities—surface energy and surface tension—are not identical, they are interchangeable for most practical purposes. The dimensions of $\sigma$ are energy/area or force/length.

The physical interpretation of surface

[1] Manuscript received December 30, 1947.

tension is, briefly, as follows: Any atom or molecule in a homogeneous phase is in equilibrium under the combined effects of the forces exerted on it by its neighbors. At the surface between phases 1 and 2 an atom is submitted to two different effects, and there is usually a resultant force directed toward the interior of one of the phases, which tends therefore to contract, reducing the area of the interface to a minimum. As the surface tension results essentially from differences in the intermolecular forces on both sides of the boundary, it will depend essentially on the chemical nature of the phases, the nature and magnitude of intermolecular forces, number of atoms within a certain distance of the boundary, i.e., the density of the two phases or, in the case of solid phases, the type of packing and dimensions of the unit cell. Surface tension will, as a rule, be a function of pressure, temperature, and composition of the phases in contact. The surface tension of a substance has no absolute value; we must still define the nature of the phase in contact with it. For crystals the surface tension and surface energy depend on the orientation of the face.

Surface phenomena at the contact of two phases are usually localized in a film of finite thickness. For thermodynamic treatment it is convenient to define an ideal surface of separation, both phases being assumed to remain homogeneous right up to this ideal surface. We assume that surface forces act in this ideal surface only and determine its position in such a manner that the ideal system will be mechanically equivalent to the actual one. It is usually necessary to assign to this ideal surface a certain concentration, called the "adsorption," and defined by

$$\Gamma_i^a = \frac{n_i^a}{a}, \tag{1}$$

where $n_i^a$ is the number of moles of $i$ in the actual system *minus* the number of moles that would be present if the two phases remained strictly homogeneous right up to the contact ($\Gamma_i^a$ may thus be positive, negative, or zero). In the same manner it is usually necessary to assign to the ideal surface a certain amount of internal energy $E^a$ and a certain entropy $S^a$, so that the total internal energy and total entropy will be the same in the ideal and the actual systems. We then have[2]

$$dE^a = T^a dS^a + \sigma\, d a + \sum_i \mu_i^a dn_i^a, \tag{2}$$

which leads to Gibbs's fundamental relation

$$S^a dT^a + a\, d\sigma + \sum_i n_i^a d\mu_i^a = 0. \tag{3}$$

Dividing by $a$, we obtain

$$s^a dT^a + d\sigma + \sum_i \Gamma_i^a d\mu_i^a = 0, \tag{4}$$

and for any change at constant temperature

$$d\sigma = -\sum_i \Gamma_i^a d\mu_i^a. \tag{5}$$

Equation (4) is quite similar to the usual relation for volume phases,

$$-s^a dT + v^a dP - \sum_i N_i^a d\mu_i^a = 0. \tag{6}$$

The conditions for equilibrium in a system of $c$-constituents and $\varphi$-phases $\alpha, \beta, \ldots, \varphi$ separated by $\varphi - 1$ interfaces $(\alpha\beta) \ldots$ are

$$\mu_i^\alpha = \mu_i^\beta = \ldots = \mu_i^{\alpha\beta} = \ldots \quad (i = 1, \ldots, c), \tag{7}$$

$$T^\alpha = T^\beta = \ldots = T^{\alpha\beta} = \ldots, \tag{8}$$

$$\left.\begin{aligned} P^\alpha - P^\beta &= \left(\frac{1}{\rho_1} + \frac{1}{\rho_2}\right)^{\alpha\beta} \sigma^{\alpha\beta}, \\ P^\alpha - P^\gamma &= \ldots, \end{aligned}\right\} \tag{9}$$

[2] For proof of the following formulas the reader is referred to one of the standard texts on the subject, e.g., Guggenheim (1934) or Defay (1934).

$\rho_1$ and $\rho_2$ being the principal radii of curvature of surface $(\alpha\beta)$; they are considered positive if the corresponding centers of curvature are on the $\alpha$-side of the surface. Putting $1/r^{\alpha\beta} = \frac{1}{2}(1/\rho_1 + 1/\rho_2)^{\alpha\beta}$, we have

$$P^{\alpha} - P^{\beta} = \frac{2\sigma^{\alpha\beta}}{r^{\alpha\beta}}, \qquad (10)$$

which may be written in the differential form,

$$dP = 2d\left(\frac{\sigma}{r}\right). \qquad (11)$$

Relation (10) is of fundamental importance. It may be put under a number of classical forms, such as the Kelvin equation, $\log p_1/p_0 = 2\sigma v/rRT$, relating the vapor pressure of a pure liquid to the radius of the drops; or the Thomson equation, $(T_0 - T_r)/T_0 = (2\sigma/r)(M/Qd)$, relating the melting-point of a grain of radius $r$ to that of a grain with plane surfaces.

For a system of $c$-constituents and $\varphi$-phases separated by $(\varphi - 1)$ interfaces, $b$ of which are plane and $(\varphi - 1 - b)$ are curved, the variance $w$ is

$$w = c + 1 - b, \qquad (12)$$

which reduces to the usual form $w = c + 2 - \varphi$ for $b = \varphi - 1$ and to

$$w = c + 1 \qquad (13)$$

for $b = 0$.

We now proceed to indicate a few geological processes in which surface phenomena may play an important part.

### SOLUBILITY-METASOMATISM

Consider a pure solid $i$, and let $\sigma$ be the surface tension at the curved interface between the pure amorphous solid and its saturated solution in any solvent. According to equation (13), the variance of the system is 3, and the state of the system will be defined by three arbitrary variables, for instances, $P$, $T$, and $r$, $r$ being the curvature of the interface. Then, at $P$, $T$ constants

$$\frac{dN_i}{dr} = \frac{-2\dfrac{\sigma \bar{v}_i}{r^2}}{\left(\dfrac{\partial \mu_i}{\partial N_i}\right)_{P,T,r}}; \qquad (14)$$

and since $(\partial\mu_i/\partial N_i)_{P,T,r} > 0$ in any stable phase, $dN_i/dr < 0$ if $\bar{v}_i > 0$. Thus the solubility increases when the grain size decreases, and small fragments are more soluble than larger ones. If small and large grains are simultaneously in contact with a given saturated solution, the smaller grains will dissolve, while precipitation occurs around the larger ones, the general tendency being toward the formation of a few large grains and the reduction of the total area of the interface solid/liquid.

For an ideal solution

$$\left(\frac{\partial \mu_i}{\partial N_i}\right)_{P,T,r} = \frac{RT}{N_i}; \qquad (15)$$

and, assuming $\sigma$ and $\bar{v}_i$ to be constant in the range considered, we obtain by integration

$$\frac{N_i^2}{N_i^1} = \exp\frac{2\sigma\bar{v}_i}{RT}\left(\frac{1}{r_2} - \frac{1}{r_1}\right), \qquad (16)$$

where $N_i^1$ and $N_i^2$ are, respectively, the molar fractions of $i$ in a solution saturated with respect to grains of radius $r_1$ and $r_2$. Take, for instance, $\sigma = 10^3$ cgs, $\bar{v}_i = 100$ cm.$^3$, $T = 300°K$, $r_1 = 10^{-5}$ cm., $r_2 = 10^{-3}$ cm.; then

$$N_i^1 = 2.22 N_i^2.$$

If the solid is in contact with a solution contained in pores of the solid itself, then the sign of the curvature $r$ is reversed, and so are the signs in equation (14). It follows that if $\bar{v}_i > 0$ the solubility will decrease in smaller openings, which will

tend to close, the larger openings becoming larger, and the general tendency being, as usual, to decrease the total area of contact.

Thus when a saturated solution of a number of constituents is introduced at constant temperature into a small pore, it will become oversaturated with respect to all constituents for which $\bar{v}_i > 0$ and undersaturated with respect to those for which $\bar{v}_i < 0$. The solution will thus generally precipitate some of its constituents and will be able, at the same time, to dissolve further amounts of some others. It follows that the behavior of a solution impregnating a rock will depend on the size of the pores and that minerals which might precipitate in a large opening may go into solution in a smaller one, or vice versa. This is a common observation (Lindgren, 1933, p. 173), for which the writer has been unable to find a correct explanation in the literature.[3]

The effect of a given solution impregnating a given rock will depend on the following factors: the magnitude of the surface tension of the solution against the various minerals in the rock; the shape and size of the cavities and pores; the shapes and sizes of individual grains in contact with the solution; and the partial molar volume of the various constituents of the solution, to which factors we must add, of course, all those which are usually considered in this connection, viz., pressure, temperature, permeability, etc.

Thus the behavior of a given solution with respect to metasomatic replacement may be expected to be extremely diversified. It is practically a hopeless task to predict, in the case of a multicomponent solution impregnating rocks of different composition and porosity, which substances will precipitate and where and when and which substances will be carried away.

### CRYSTALLIZATION

Because a small grain is more soluble than a large one, it follows that, when precipitation or crystallization begins, i.e., when the dimensions of the incipient grains are extremely small, there must be a certain amount of supersaturation with respect to a larger mass of the solid phase. This is why crystallization starts more readily when "germs" of finite size are present. It has been shown statistically (Landau and Lifshitz, 1938) that the probability of formation of a germ of given radius is a function of $e^{-\sigma}$ and thus decreases rapidly when the surface-tension grain/solution increases; in some cases, germs may have to be added from outside.

The problem becomes more complicated in the case of crystals, because $\sigma$ will then be different in different directions. Gibbs (1928, p. 320) has shown that the condition for stability, for a given volume $v$ of the crystal, is that $\sum_i a_i\sigma_i$ be a minimum, where $a_i$ is the area of a face with surface energy $\sigma_i$, the summation being extended to all faces of the crystal. Gibbs's equation (666) states more explicitly that the quantity

$$\frac{\Sigma\,(\sigma'_i l_i \operatorname{cosec} \omega_i - \sigma_i l_i \cot \omega_i)}{a_i} \quad (17)$$

[3] Bain (1936) explains such facts by the variation of pressure which, according to Bernoulli's theorem, must result from variations in the velocity of the solution. Now this pressure variation is $\delta P = \frac{1}{2}\rho(\delta v)^2$ where $\rho$ is the density of the solution and $\delta v$ is the change in the velocity. Since the velocity of the solution probably never exceeds a few centimeters per second, $\delta v$ is less than this, and, accordingly, the change in chemical potential of constituent $i$ $\delta\mu_i = \bar{v}_i\delta P$ is only of the order of $10^2$ ergs for $\bar{v}_i = 100$ cm³. From equation (11), on the contrary, we find for $\sigma = 10^3$ cgs, $r = 10^{-5}$ cm., variations of $\mu_i$ which are of the order of $10^8$ ergs. The effects of changes in velocity are thus quite negligible compared to surface-tension effects.

must be the same for all faces, $a_i$ being the area of the face on which the crystal grows, $l_i$ the length of the common edge between this face and an adjacent face of area $a_i'$ and surface energy $\sigma_i'$, the summation being extended to all faces adjacent to face $i$. The quantity $\omega_i$ is the external angle between the two faces. The expression (17) measures the differences in chemical potential between face $i$ and the same amorphous solid. As Gibbs points out, the value of the chemical potential in the solution which will be necessary for the growth of the crystal face will usually be different from that required for precipitation of the same solid in the amorphous state and will generally be greatest for the faces for which $\sigma$ is least.

It has been shown (Volmer, 1939) that if, for example, we start from a prismatic crystal having faces 1, 2, . . . , and measure the growth normal to each face by the lengths $h_1$, $h_2$, . . . , we must have, at equilibrium,

$$\frac{\sigma_1}{h_1}=\frac{\sigma_2}{h_2}=\cdots, \tag{18}$$

a relation known as "Wulff's law." The growth of an incipient crystal is therefore essentially determined by the surface energy of its various faces.

## RECRYSTALLIZATION

As has been pointed out, there will always be a tendency for any system under given conditions to reduce the area of the interfaces to a minimum. This tendency corresponds to the tendency for a rock to become equigranular, smaller crystals going into solution at points of maximum curvature and precipitation occurring on larger grains at points of less curvature. The presence of a solvent is, of course, not necessary; transport of material might occur by diffusion; but, as diffusion is an extremely slow process at ordinary temperatures, the presence of a solvent will notably accelerate the process. This type of recrystallization is quite independent of any exterior action such as heat, pressure, or stress, and it may occur before, during, or after any mechanical deformation to which the rock may be submitted. In the same manner, disseminated amounts of a given mineral will tend to concentrate in larger units (porphyroblasts). This tendency to form porphyroblasts will depend essentially on the magnitude of the surface energy which, as we have mentioned, is a function essentially of the magnitude of intermolecular forces, type, and closeness of packing. Surface energy may thus be an important factor in metamorphic processes.

## VAPOR PRESSURE

The composition of a gas phase in equilibrium with a magma will depend, because of equation (10), on the size of the bubbles. Two bubbles of the same composition but of different size cannot simultaneously be in equilibrium. Bubbles of different sizes must necessarily have different compositions.

In the same manner, the composition of the gas phase permeating the pores of the wall rock will be different from that of a gas phase in equilibrium with the magma through a plane interface.

Conversely, the composition of the gases held in solution in, or adsorbed on, a crystal is not necessarily the same as that of the gas phase that was being given off by the magma at the time of crystallization. Shepherd (1938) has already called attention to the fact that adsorption may complicate in a hopeless manner the determination of the exact composition of a magmatic gas phase.

It is interesting that the vapor pres-

sure of a magma may be determined from adsorption and surface-tension data. Consider, for instance, a simple system consisting of a pure gas (phase 1, constituent 1), a solution (phase 2, constituents 1 and 2), and a pure solid (phase 3, constituent 2). Such a system would be normally univariant. If the interfaces are curved, the variance is 3. The equations are (*a*) three equations of type (6), one for each phase; (*b*) two equations of type (10); and (*c*) two equations of type (4). These may be written as follows:

$$\left.\begin{aligned}
&-v^1dP^1+s^1dT+d\mu_1=0\ ,\\
&-v^2dP^2+s^2dT+N_1d\mu_1+N_2d\mu_2=0\ ,\\
&-v^3dP^3+s^3dT+d\mu_2=0\ ,\\
&dP^1-dP^2=\frac{2}{r^{12}}\,d\sigma^{12},\\
&dP^2-dP^3=\frac{2}{r^{23}}\,d\sigma^{23},\\
&-s^{12}dT-d\sigma^{12}-\Gamma_1^{12}d\mu_1-\Gamma_2^{12}d\mu_2=0\ ,\\
&-s^{23}dT-d\sigma^{23}-\Gamma_1^{23}d\mu_1-\Gamma_2^{23}d\mu_2=0\ .
\end{aligned}\right\}\tag{19}$$

First, suppose that all the adsorptions are negligible. Eliminating $d\sigma^{12}$, $d\sigma^{23}$, $dP^2$, $dP^3$, $d\mu_1$, and $d\mu_2$, we find for the variation with temperature of the vapor pressure $P^1$

$$\frac{dP^1}{dT}=\frac{N_1\Delta s_1-N_2\Delta s_2-N_2v^3\left(2\,\frac{s^{23}}{r^{23}}+2\,\frac{s^{12}}{r^{12}}\right)+v^2\,\frac{2\,s^{12}}{r^{12}}}{N_1\Delta v_1-N_2\Delta v_2}\,,\tag{20}$$

where

$$\Delta v_1=v^1-\bar v_1^2\,,\qquad \Delta S_1=s^1-\bar s_1^2\,,$$
$$\Delta v_2=\bar v_2^2-v^3\,,\qquad \Delta S_2=\bar s_2^2-s^3\,.$$

For $r^{12}=r^{23}=\infty$, this equation reduces to the familiar relation for the vapor pressure of a univariant system of two constituents (Goranson, 1938, p. 84).

Now take $r^{23}=\infty$, $\Gamma_1^{23}=\Gamma_2^{23}=\Gamma_2^{12}=0$, $\Gamma_1^{12}\neq 0$. We then have

$$\frac{dP^1}{dT}=\frac{N_1\Delta s_1-N_2\Delta s_2+\frac{2}{r^{12}}\,(s^{12}-\Gamma_1^{12}s^1)\,(N_1\bar v_1^2+N_2\Delta v_2)}{N_1\Delta v_1-N_2\Delta v_2-\frac{2\,v^1}{r^{12}}\,\Gamma^{12}\,(N_1\bar v_1^2+N_2\Delta v_2)}\,.\tag{21}$$

The quantity $s^1$ is usually much greater than $\Delta s_1$, and $\bar v_1^2$ for water vapor is of the order of several tens of cubic centimeters in the temperature-pressure range considered, so that $(2/r^{12})\Gamma_1^{12}s^1N_1\bar v_1^2$ may be of the same order of magnitude as $N_1\Delta s_1$ if $\Gamma_1^{12}/r^{12}$ is $10^{-2}$ or greater. The same applies to the last term in the denominator, which may become of the same order as the first time. Adsorption may, therefore, introduce an appreciable correction in the vapor-pressure curve.

Finally, if $r^{12} = r^{23} = \infty$ but $\Gamma_1^{12}$ and $\Gamma_2^{23} \neq 0$, the equations are, omitting unnecessary indices,

$$\left.\begin{aligned} &-v^1dP+s^1dT+d\mu_1=0\ , \\ &-v^2dP+s^2dT+N_1d\mu_1+N_2d\mu_2=0\ , \\ &-v^3dP+s^3dT+d\mu_2=0\ , \\ &-s^{12}dT-d\sigma-\Gamma_1d\mu_1-\Gamma_2d\mu_2=0\ . \end{aligned}\right\}(22)$$

The first three equations yield the usual relation

$$\frac{dP}{dT}=\frac{N_1\Delta s_1 - N_2\Delta s_2}{N_1\Delta v_1 - N_2\Delta v_2}, \qquad (23)$$

whereas, by eliminating $d\mu_1$ and $d\mu_2$, we obtain

$$\frac{dP}{dT}=\frac{\Gamma_1 s^1+\Gamma_2 s^2 - s^{12}+\dfrac{d\sigma}{dT}}{\Gamma_1 v^1+\Gamma_2 v^2}. \qquad (24)$$

Equations (23) and (24) are thus equivalent, and either may be used. Equation (23) is, of course, the easier one to use, as quantities such as adsorptions and surface entropies are difficult to measure. However, the vapor pressure of this simple system *could* be calculated from surface-tension data alone.

## ADSORPTION

Let us consider a rock containing a certain amount of water adsorbed on its surface and compute roughly the pressure under which this water would be liberated if the adsorption forces ceased suddenly to exist. Let $A$ be the surface per unit mass of the rock ($A$, for some substances, may be of the order of $10^7$ cm²/gr). Suppose that all the pores are of radius $r$. Then the number of pores, $n$, is

$$n=\frac{A}{4\pi r^2},$$

and their total volume is $\frac{1}{3}Ar$. If $n^\alpha$ is the number of moles of a perfect gas adsorbed in the pores, the pressure that would develop in the pores if the adsorption forces suddenly ceased to exist would be

$$P^\alpha = n^\alpha \frac{3RT}{A\,r}.$$

Hence

$$n^\alpha = P^\alpha \frac{A\,r}{3RT}.$$

The number of moles $n^\beta$ that would be present in the pores at the outside pressure $P^\beta$ is

$$n^\beta = P^\beta \frac{A\,r}{3RT};$$

and the adsorption $\Gamma$ is therefore

$$\Gamma=\frac{n^\alpha - n^\beta}{A}=\frac{\Gamma}{3RT}(P^\alpha - P^\beta)\ ; \quad (25)$$

and, finally,

$$P^\alpha = \frac{3RT\Gamma}{r}+P^\beta. \qquad (26)$$

Consider a rock with pores of radius $10^{-5}$ cm., surface $10^3$cm²/gr, and containing 0.18 per cent of adsorbed water, so that $n^\alpha = 10^{-4}$ moles/gr. The adsorption $\Gamma$ is then $10^{-4}/10^3 = 10^{-7}$, and, at 1,000° C.,

$$P^\alpha = 3{,}240 + P^\beta \qquad \text{(bars)}.$$

Adsorption is known to decrease rapidly when temperature increases. It follows that if a rock containing as little as 0.18 per cent of adsorbed water could be heated up rapidly to 1,000°—a temperature at which we assume that desorption is complete—the water vapor would be released under very high pressure. The figure obtained is, of course, only approximate, because at this pressure the gas ceases to be perfect and, moreover, the postulate of spherical porosity is probably not satisfied. Nonetheless, very high pressures may originate in this manner, and desorption resulting

from rapid heating may be a factor in some types of explosive volcanic activity.

It is believed also, although the matter has not been thoroughly investigated, that surface tension may control to some extent the surficial aspects of lava flows. Surface tension, as is well known, controls the spreading of a liquid on a solid surface and also the size and shape of drops formed when a thin vein of viscous liquid flows through a small opening. The size of the bubbles which may form in a lava flow and the growth of crystals in this lava are, as we have seen, also controlled to some extent by surface tension. These factors are probably of some importance in determining the surficial aspects of lava flows.

It must be emphasized, however, that surface-tension effects such as we have considered depend essentially on the curvature of the interfaces and become vanishingly small if the curvatures become small.

## REFERENCES CITED


BAIN, G. W. (1936) Mechanics of metasomatism: Econ. Geology, vol. 31, pp. 505–526.

DEFAY, R. (1934) Étude thermodynamique de la tension superficielle, Paris, Gauthier-Villars.

GIBBS, J. W. (1928) The collected papers of J. Willard Gibbs, New York, Longmans, Green & Co.

GORANSON, R. W. (1938) Silicate-water systems: phase equilibria in the $NaAlSi_3O_8$–$H_2O$ and $KAlSi_3O_8$–$H_2O$ systems at high temperatures and pressures: Am. Jour. Sci., vol. 35A, pp. 71–91.

GUGGENHEIM, A. E. (1934) Modern thermodynamics by the method of Willard Gibbs, London, Methuen.

LANDAU, L., and LIFSHITZ, E. (1938) Statistical physics, Oxford University Press.

LINDGREN, W. (1933) Mineral deposits, New York, McGraw-Hill Book Co.

SHEPHERD, E. S. (1938) The gases in rocks and some related problems: Am. Jour. Sci., vol. 35A, pp. 311–351.

VOLMER, MAX (1939) Kinetik der Phasenbildung, Leipzig, Steinkopff.

# HOLLOW FERRUGINOUS CONCRETIONS IN SOUTH CAROLINA[1]

LAURENCE L. SMITH
University of South Carolina

ABSTRACT

Ferruginous concretions are locally abundant in sands and sandy clays of the South Carolina Coastal Plain. The great variety of shapes is primarily determined by the structure of the enclosing sediments. Hollow forms result from intergranular deposition of limonite in excess of that necessary to fill interstices between sand grains. The outer zone of solid concretions receives greater increments of iron oxides and, being thus forced to expand more rapidly, results in hollow interiors. The rate of supply of material is an important factor in determining the form of the concretions.

## INTRODUCTION

Concretions similar to those in South Carolina have been described from other localities and are apparently of widespread occurrence, particularly in warm, humid climates. Wherever concretions are abundant, gradation from solid to hollow types is usually found. This gradation suggests that the hollow ones form by the same processes that are responsible for the solid types.

No evidence has been found of hollow interiors formed by solution of carbonate cores. In some, limonite was deposited around clay balls, which, by dehydration, formed interior cavities. But the function of the clay interior has been primarily that of a nucleus around which precipitation of the limonite was localized. In general, the hollow concretions were formed by continued deposition of limonite cement in sufficient amounts to wedge apart the sand grains.

## OCCURRENCE AND ASSOCIATIONS

Ferruginous concretions in South Carolina occur in sediments which vary from coarse sands to sandy kaolins and range in age from Cretaceous to possibly Pleistocene. They are locally abundant where ferruginous sands overlie more impervious materials, as in the Fall Line belt, where a blanket of red to buff sands rests upon clayey members of the Tuscaloosa or upon the crystalline rocks. Ferruginous fossil wood has been found in abundance at several localities in South Carolina under the latter stratigraphic conditions. The water table is also a favorable zone of limonite deposition.

The blanket of post-Cretaceous sands has a maximum thickness of about 50 feet. These sands are cross-bedded and show considerable variation in texture and color. Some consist almost entirely of quartz grains with varying amounts of limonitic cement. Some mica and opaque minerals may be present.

Mottling is generally pronounced. Above the water table this feature is due to leaching of iron, and the spots are gray or buff against a red background. Below the water table the mottling is due to deposition of limonite, the spots being various shades of red and brown, developed in lighter-colored sands. Numerous chemical analyses have shown that the red spots contain from three to seven times as much iron oxide (calculated as $Fe_2O_3$) as do the enclosing sands. The red spots generally range from 0.4 to 3 per cent iron oxide and the normal sands from 0.1 to 0.6 per cent. Thus the red

[1] Manuscript received May 20, 1947.

mottling is not due simply to oxidation of iron but to actual concentration of iron oxide.

The sands beneath surface depressions, in which organic remains have accumulated, are uniformly bleached and leached of iron, which is carried to lower levels. Sections exposing the former position of the water table reveal stratiform ferruginous deposits at this level and also just above relatively impervious layers. The writer has found similar conditions in drilling undrained depressions. From this, together with the widespread mottling, it is obvious that in warm, humid climates iron is readily leached, transported, and redeposited in sediments. Gradation from limonite-rich spots to solid and hollow concretions has been found, and all result from the same general processes.

## COMPOSITION AND FORM OF THE CONCRETIONS

The concretions are composed of the same materials as the enclosing sediments, differing only in the greater percentage of iron oxide cement.

Stratiform concretionary deposits are the most common type. Thin layers, commonly interbedded with the sands, are localized just above relatively impervious bases. Most kaolin beds are capped by iron-cemented layers, in places several feet thick and of sufficient firmness to serve as dimension stone. Scattered concretions of all shapes may occur just above the stratiform deposits or may be intergrown with them in such a manner as to indicate that the smaller concretions developed first and were later incorporated in the limonite-cemented layer.

Isolated concretions are variously spheroidal, ellipsoidal, lens-shaped, fusiform, rod-shaped, cylindrical, or nodular. They range in size from a fraction to several inches in thickness, and tubular forms 3 feet long have been found. Smaller concretions are most likely to be spheroidal; larger concretions are less regular. One shape or related shapes characterize a given zone.

Scattered concretions exhibit great variation in size of interior cavities, which bear no relation to the size of the whole concretion. Large concretions may have incipient or small cavities surrounded by thick walls, whereas many small ones may have thin shells, with cavities amounting to 75 per cent of the volume of the whole. The shapes of the cavities conform closely to the outer form, elongated concretions having tubular openings and flat forms containing voids of comparable outlines. The shells are generally quite uniform in thickness, particularly in the spheroidal forms. Unsymmetrical types, with one side more highly arched, show the more convex wall to be thicker.

The interiors are partially filled with ferruginous sand, which in most cases differs from the enclosing shell only in compactness and amount of limonite cement. Only in concretions formed around clay balls is there any difference between the shell and the core material. Some cores are compact and show only slight separation from the walls; others consist of vuggy, weakly cemented sands; and many have hollows partially filled with loose sand, which rattles when the concretion is shaken.

Chemical analyses show that iron oxide constitutes a larger proportion of the shell material than of the loose interiors. Examinations of thin sections showed that the percentage of ferruginous cement is greatest at approximately the middle of the shells.

Concentric fractures develop in con-

cretions with incipient hollows (pl. 1). The fractures conform to the surface of the concretions and are more or less parallel to one another. Drusy limonite coatings indicate that the cracks formed before completion of limonite deposition.

Microscropic examination of thin sections shows the sand grains widely separated by intergranular limonite cement (fig. 1). The mica flakes are bent and the quartz shattered by forces attendant upon deposition, the amount of disruption increasing with increase in amount of limonite.

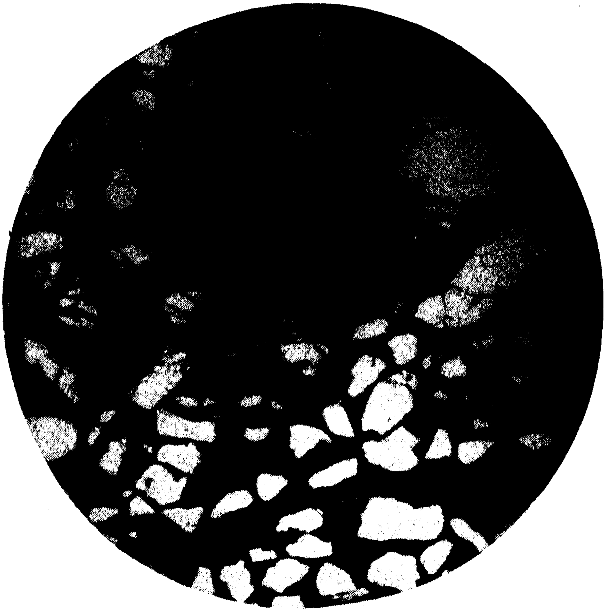


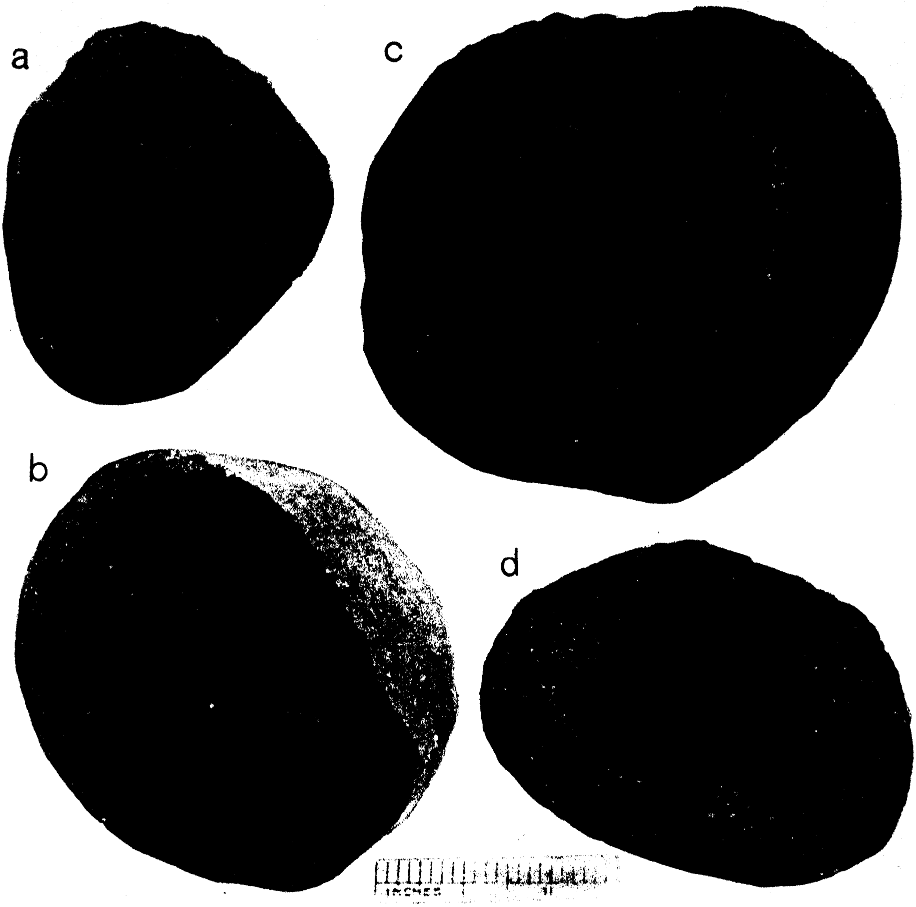
FIG. 1.—Photomicrograph showing section of shell of hollow ferruginous concretion. All of black area is limonite. Intergranular deposition of limonite has bent the mica and shattered quartz grains. ×35.

## ORIGIN OF HOLLOW FERRUGINOUS CONCRETIONS

Iron, leached from ferruginous sands by meteoric waters, is probably transported in the form of ferric oxide hydrosol. The more thorough leaching beneath depressions containing organic matter indicates that solution is facilitated by organic acids. Deposition at depth begins around some local precipitating agent, and cementation of the sands takes place by interstitial secretion of limonite. Deposition proceeds outward from the initial locus, and, because the building material is received from the surrounding area, the periphery of the concretion grows most rapidly in directions of greatest supply, which accordingly determines the shape.

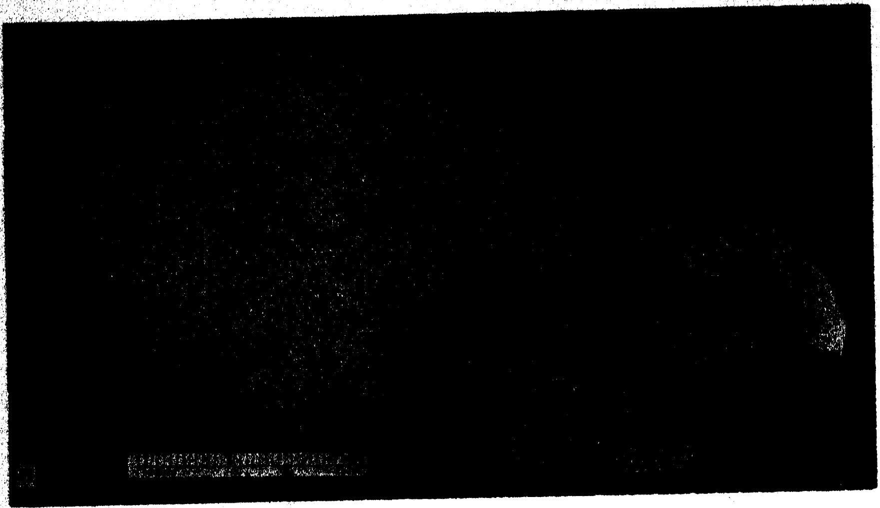
After interstitial spaces are filled, additional deposition of limonite forcefully wedges the sand grains apart. The crustal area expands, whereas the core receives little or no secretion after completion of interstitial filling, with the result that the growing concretion develops a hollow interior.

PLATE I

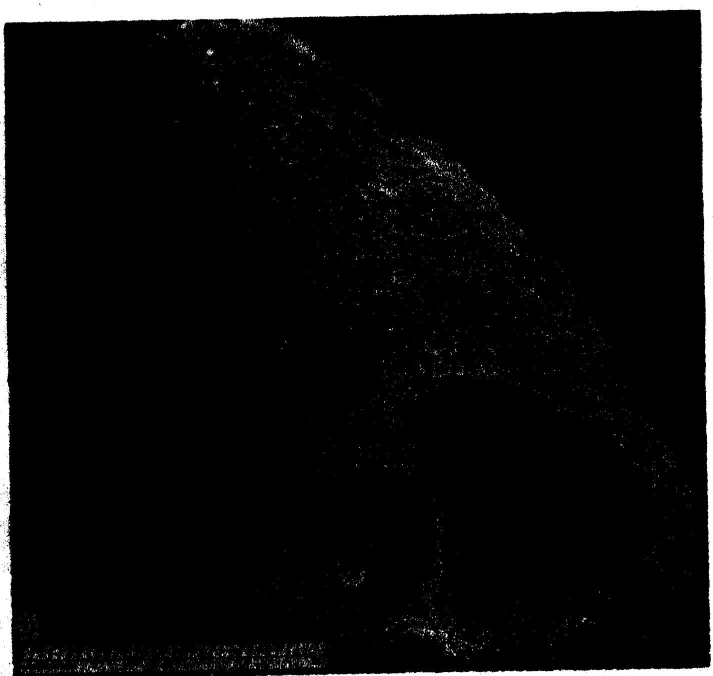



Ferruginous concretions from near Patrick, S.C., showing stages in development from solid into hollow forms. Note concentric tension fractures in *a* and *b*, separation of core in *b* and *c*, and loose sand and iron oxide in interior of *d*.

PLATE II



A



B

*a*, Ellipsoidal hollow ferruginous concretion. The interior is partially filled with weakly cemented sand. *b*, Bicameral tubular ferruginous concretion. Note greater thickness of outside wall of small chamber.

Todd (1903, pp. 353–368), noting calcareous concretions with cracks radiating from the center, postulated that the outer portion had

> expanded and become too large for the interior, and has done so with sufficient force to wrench apart the interior. Solidification takes place . . . largely between the particles already deposited, wedging them apart with force sufficient to separate the portion inside the expanding zone.

He called such growths "intercretions" and, to the writer's knowledge, is alone in offering such an explanation for any type of concretion. However, hollow limonitic concretions which do not show comparable radiating cracks were explained by Todd (1903) as due to deposition of iron oxide around a ferrous carbonate "intercretion," with subsequent solution of the included core.

Bates (1938), in discussing hollow limonite concretions occurring in certain tills of Iowa, thinks that Todd's explanation seems applicable. He says:

> The history of hollow concretions may be summarized as follows: (1) the formation in glacial till of siderite concretions most of which were not pure siderite but included some of the surrounding material . . . (2) solution of the iron carbonate and precipitation of ferric hydroxide as a crust around the siderite.

Shaw (1917) offers a similar leaching process to explain hollow pebbles of iron oxide occuring in the bluff of the Mississippi River at Natchez. He states that iron compounds seem partly to have replaced the lime carbonate of limestone pebbles and that each concretion contains a little clay that is indistinguishable from the residue of limestone after the lime carbonate has been dissolved with acid. Shaw shows excellent photographs of a variety of ferruginous concretions; all his forms may be duplicated from localities in South Carolina. Concretions from Shaw's[2] locality, which the writer examined, were apparently formed by deposition around clay balls. Subsequent growth by expansion of the limonite shell, together with possible shrinkage of the clay core, left a void between crust and interior kernel. A similar explanation is believed to apply to the hollow limonite concretions described by Bates.

Tarr (1935), in his excellent treatise on concretions in the Connecticut Valley, believes that the cementing material does not exceed the available pore space. Concerning the calcareous concretions, which are far more numerous in that locality than are the ferruginous ones, he regards the 35–53 per cent of $CaCO_3$ cement as significant, since this amount fits rather closely to the maximum porosity of the enclosing silt. He says:

> The introduction of more than 50 per cent of $CaCO_3$ would have resulted either in a disturbance of the beds about the concretions or in replacement . . . concretions formed in disturbed beds preserve in perfect detail the folds and faults of the enclosing layers. The concretions are simply cemented portions of the enclosing layers.

It is to be noted that these concretions are very impure, the calcareous types containing considerable magnesium carbonate, iron, and aluminum oxides. These impurities suggest rapid deposition, as Tarr concluded from other evidence, and growth would proceed by enmeshing addition silt, with no tendency for intergranular secretion beyond that necessary to fill the voids. The writer has invariably found relatively large cavities within ferruginous concretions formed in kaolins. Here growth has been particularly slow and constant, with opportunity for the diffusion of molecules to the supersaturated adsorbed films in contact

[2] Specimens furnished Stephen Taber through courtesy of Dr. Shaw.

with interstitial limonite. Thus deposition tends to continue adding to that already present, and intergranular secretion would dominate, whereas rapid deposition would create new centers of precipitation and enmeshing of additional particles would result.

Although Tarr (1935) finds sedimentary structures preserved and no evidence of concretions making room for themselves, others (Daly, 1900, and Taber, 1919) have offered graphic evidence of arching of bedding planes by growth of calcareous concretions.

In hollow concretions examined by the writer, iron oxide (calculated as $Fe_2O_3$) constitutes over 60 per cent of the shells, whereas the loose core material contains less than 50 per cent. The intergranular limonite tends to obliterate all primary sedimentary structures, which are, therefore, indistinguishable inside the oxide shell. In solid forms bedding may be traced through the concretion.

Growth during the early stages of solid ferruginous concretions proceeds rapidly and largely by enmeshing of particles. Thus arching of enclosing materials does not take place. The concretions of other writers, which show displacement of beds, have grown slowly with consequent exclusion of surrounding particles. Evidence that expansion of hollow types arches the enclosing materials has not been found in South Carolina and would not be expected because of the loose character of the sediments. However, the exertion of effective forces by intergranular deposition is convincingly shown by the shattered quartz grains and bent mica flakes (fig. 1), as well as by the concentric fractures, which intersect cement and included grains alike. The disruption of minerals increases with greater amounts of limonite and is absent where cementing material is only sufficient to fill the interstices. Drusy limonite coatings on the fracture surfaces show that breaking took place during the process of secretion.

### LOCI AND MODE OF DEPOSITION

Initial precipitation is most commonly localized by some type of void, such as spots of coarser sand, minute cavities, and, to some extent, root holes and fractures. Deposition also begins around clay balls, fossil wood, and possibly pre-existing particles of iron oxide.

Sands which have become mottled by leaching are depleted of iron from existing voids, such as coarser-textured spots and fractures. Below the water table, where mottling is due to deposition, the voids serve as locales of concentration. Limonite is first precipitated upon the surfaces of coarser particles,—filling of interstices follows, and, finally, the mass is cemented into a concretion. Alternate wetting and drying may explain initial precipitation in some voids above the water table, but within the zone of saturation the principal factor is believed to be the greater mobility of dispersed sol molecules where voids are larger. Supersaturation is first attained in the layers of liquid held to the walls of particles by adsorption and would be attained simultaneously throughout the whole system. However, the greater volume of solution in the larger voids permits more effective diffusion of molecules, and so centers of deposition are selectively developed within existing cavities. Once precipitation is begun, equilibrium is maintained by diffusion of colloidal particles from the surrounding zone, and deposition continues by molecular attraction as long as limonite and saturated solution are in contact.

Small spherical cavities, probably due to solution and common in kaolinitic

clays, are conspicuous points of initial precipitation. Their surfaces are first coated with limonite, which either thickens into an incrustation or forms a meshlike aggregate within the opening. Later secretion extends outward very slowly, resulting in maximum intergranular deposition, as explained above, and thus shells are readily expanded, with interior cavities soon becoming relatively large. Spherical tension cracks dividing the walls into concentric layers attest to the efficacy of the expansion of the outer shells.

Shrinkage cracks and root holes may be incrusted with limonite, but neither are important in localizing deposition, probably because they are not common below the water table. There is no evidence that thick-walled tubular concretions have grown inside root cavities.

Many concretions develop around clay balls or clods of sand weakly held together by clay. These are grouped along definite planes, where the balls were probably formed by waves. The balls, being less pervious to large diffusing molecules of ferric hydroxide, selectively filter them out around their periphery. After definite crusts are formed, continued intergranular secretion results in expansion, and the included ball becomes loose and only partially fills the interior. The hollow concretions described by Shaw (1917) have this origin, and several of his specimens examined by the writer contain a loose kernel of hardened homogenous clay. Concretions that have formed around a ball of partially cemented sand now contain a loose mass of rattling particles left by the drying of the adhering clay.

Fragments of fossil wood were found inside two medium-sized spheroidal concretions. They were attached to one side of the inner wall, and the cavity was partly filled with a residue of sand and limonite particles. The wood apparently served to localize deposition in the same manner as did the clay balls.

### FEATURES CONTROLLING GROWTH AND SHAPES OF CONCRETIONS

The various shapes of concretions and ferruginous aggregates have been determined by the structures of the enclosing sediments.

Stratiform concretionary deposits occur above relatively impermeable layers. Probably such limonitic beds are formed just as readily above as below the water table, provided that there is a sufficient overburden of ferruginous material from which iron may be leached.

Spheroidal types occur in relatively structureless materials ranging from coarse sands to kaolins. In such sediments iron molecules are transferred toward the growing aggregates with equal ease in all directions, and additions are deposited symmetrically.

Compound spheroidal or nodular concretions are generally found in association with simple spheroidal types. Some are the result of attachment of additional units to earlier-formed aggregates, and the nodules simply have one wall in common. Others are composed of hollow nodules, the chambers of which are mutually confluent, indicating that the individuals merged early in the process of accretion and that all developed cavities concurrently with their later growth.

Spheroidal concretions and stratiform deposits represent the extremes between which there is a wide range of variations. Rounded types predominate in homogenous sediments and flat ones in materials with marked stratification. In a kaolin pit near Edmund, South Carolina, an abundance of spheroidal concretions occurs in the upper part of massive-

appearing clay. There are also some pod- and lens-shaped types which have their longer axes horizontal and obviously concordant with obscure bedding. The contact of the kaolin with an overburden of sands is marked by limonite-cemented sheets. Associated with these is an assortment of flat concretions but no spheroidal types.

A highway cut, 8 miles south of Lugoff, South Carolina, exposes ferruginous sands with pronounced horizontal structure. Limonite layers 1 inch in thickness parallel the bedding and spread out horizontally for several feet. Along the same horizons are numerous detached concretions which have flat shapes simulating lenses, ellipsoids, or pods (pl. 2, *A*). Many of them are scarcely 1 inch thick while measuring 1 foot horizontally, yet all that were examined contained cavities. Apparently, the stratiform aggregates were supplied almost entirely from above, while detached flat concretions probably received additions from below and laterally as well as by downward diffusion.

Tubular concretions are of two kinds: slender detached ones, which are extremely elongated spindles, and others formed by the separation and arching of limonitic layers along an axis.

Most of the detached tubes are the size of fountain pens, but they have been found up to nearly 3 feet in length and 4 inches in diameter. The length of most large specimens is indeterminable, since only segments are recovered from exposures in embankments. These segments give a false appearance that the tubes are open at both ends, whereas entire specimens always terminate in solid tapering cones and the interiors contain loose aggregates of sand and limonite. The diffusing molecules were fed toward these tubes from opposite directions and normal to the length, which always conforms to the bedding. A partially cemented rodlike aggregate is first developed, later an outer shell, and, finally, a tube results from excessive intergranular secretion and expansion of the wall.

Expansion and separation of crusts from limonitic layers produces planar cavities (pl. 2, *A*) or tubular openings if the arching is definitely linear. Such tubes may be variously angular in section or cylindrical, depending upon the perfection of arching. Numerous tubes with long slitlike, angular, or distinctly cylindrical openings have been found along the upper surface of a large ferruginous bed in the Fort Jackson Reservation near Columbia, South Carolina. Gradations from incipient slitlike cavities to tubular ones are common. Later thickening of the limonitic bed has incorporated many tubes within its mass. Obviously, these concretions grew largely by additions from above, although during the arching process some lateral transfer may have occurred.

Compound tubular concretions originate after the manner of some nodular types except that structure modifies the diffusion of building material in the former. The pictured specimen (pl. 2, *B*) must have first acquired the smaller chamber (note the thinner wall), and the larger chamber later formed beside it. The outer wall of the smaller represents the thickness of the original tube plus additions made by the later one. Rates of growth were not uniform throughout the whole length, for cross-sections vary in size and outline, with numerous corrugations and nodules. The secondary chamber is believed to have resulted from the speeding-up of deposition along one side of the original tube. A new rod-

like aggregate resulted from relatively rapid enmeshing of sand grains; and later with slower deposition it acquired a surrounding wall to be followed by an inner cavity similar to that of the first tube. A repetition of this process would result in polychambered forms. Willcox (1914) has described the published excellent photographs of such polychambered tubular concretions from the Redbank sands in New Jersey. These, like the ones from South Carolina, can be satisfactorily explained by the process of intergranular secretion of limonite.

## REFERENCES CITED

Bates, R. L. (1938) Occurrence and origin of certain limonite concretions: Jour. Sedimentary Petrology, vol. 8, no. 3, pp. 91–99.

Daly, R. A. (1900) The calcareous concretions of Kettle Point, Lambton County, Ontario: Jour. Geology, vol. 8, pp. 135–150.

Shaw, E. W. (1917) The Pliocene history of northern and central Mississippi: U.S. Geol. Survey Prof. Paper 108, pp. 125–163.

Taber, Stephen (1919) The mechanics of vein formation: Am. Inst. Min. Met. Eng. Trans., vol. 61, pp. 3–36.

Tarr, W. A. (1935) Concretions in the Champlain formation of the Connecticut Valley: Geol. Soc. America Bull. 46, pp. 1493–1534.

Todd, J. E. (1903) Concretions and their geologic effects: Geol. Soc. America Bull. 14, pp. 353–368.

Willcox, O. W. (1914) Iron concretions of the Redbank Sands: Jour. Geology, vol. 14, pp. 243–252.

# GEOLOGICAL NOTES

## UNUSUAL VOLCANIC DIKE AND GROOVED LAVA AT AUCKLAND, NEW ZEALAND[1]

J. A. BARTRUM AND E. J. SEARLE
Auckland University College

### HOLLOW DIKE

Quarrying operations in the well-bedded scoriae of the cinder cone of Mount Wellington (448 ft.), Auckland, have, since 1926, exposed a nearly vertical dike, which shows several unusual features.

This dike, which in most places is not over 1 foot wide, is characterized by small, irregular, cupola-like extensions. The extensions, which appeared as excavation proceeded, are now, in large part, destroyed. Short apophyses are rare (fig. 1). Near the base of the cone, the dike appears to have been the source of a small flow, a little over 3 feet in maximum depth. The flow is exposed for about 15 yards in a cut, though its full extent is unknown on account of debris. The dike has been traced to within 3 or 4 yards of the flow and almost certainly joins it below the floor of the quarry. From the quarry floor it extends upward for a little over 100 feet. Its upper portions, especially near the cupolas, have a median void, 5 inches in maximum width, but diminishing in a few feet to nothing (fig. 1). On either side of this central cavity there is a wall of solid basalt about 4 inches in width in contact with the scoriae walls; this forms an arched hood over the opening at any cupola (text fig. 1). The inner surface of this wall very commonly shows signs of notable refusion, for in many places it is glazed and coated by drip (pl. 1, fig. 2).

At one place, a later intrusion of basalt, 5 inches in width, has been pushed up in the void of the dike; all that is visible of it at present is a small upward-projecting tongue with typical black glossy chilled skin over its strongly arched surface and on its sides where they make contact with the containing walls of the main dike.

Three curious spiracle-tubes have also been found associated with the dike. One occurred *in situ* immediately above the dike, whereas the other two had obviously fallen from similar situations. The largest is 2 feet, 10 inches in length and 1 foot in maximum diameter at one end and 4 by 1½ inches at the other (pl. 1, figs. 1, 3). These tubes were probably upward extensions of the cupolas.

The origin of these spiracle-tubes is closely bound up with that of the dike. The basaltic magma that was injected into the dike fissure must have been very highly charged below with occluded gas, which, on being freed, converted upper portions of the invading magma into a frothy mixture of gas and liquid. At several points there was an excape of this gas from the upper surface of the dike. Upward movement of the froth of liquid and gas was accompanied by freezing of about 4 inches of basalt to either wall of the fissure. The gases liberated on freezing escaped into the median hollow of the dike, where their combustion developed sufficient heat to fuse the walls of the cavity. The gases, on leaving the dike, fluxed tubular passages through the scoriae for a few feet and welded material adjacent to these passages to irregular cylindrical bodies.

### GROOVED LAVA

Three examples of lava with grooves very similar to those described (1938) and by Nichols and Stearns (19 from New Mexico and Idaho, have recen

[1] Manuscript received October 7, 1947.

been noted at Auckland. One is a block of basalt which has been erupted with scoriae at the Mount Wellington quarry mentioned above; another is a slab of basalt in a lava cavern in the grounds of Auckland Grammar School, Mount Eden; and the third is part of the roof of a cavern in a flow from Mount Eden at Mortimer Pass, Newmarket. The only comparable example of this grooving previously noted from Auckland has been described from a dike in a quarry also adjacent to Mount Eden (Bartrum, 1928).

Nothing is known of the conditions of origin of the Mount Wellington block of grooved lava (pl. 1, fig. 4). A mass of scoriaceous lava was so firmly welded to its grooved surface that, when a cold chisel was used to prise the two portions apart, rupture occurred through the material of the upper block, as indicated in the photograph, and not along the surface of contact. It is evident that, after the formation of the grooves, a mass of hot ejected lava fell upon the grooved surface and had sufficient liquidity to be welded firmly to it. A second small adhering fragment is shown in the left middle portion of plate 1, figure 4.

The grooved lava at Auckland Grammar School appears in a slab 5 by 1½ feet in size. The slab, completely covered by grooves or striations, has fallen from a side wall of a cavern about 12 feet across. The grooved surface is continued in solid lava at the end of the cavern in a fissure with a 45° dip. The grooves have much less depth and are much closer together than those of plate 1, figure 4, so that many are better regarded as striations. The surface of the slab is roughened by close-spaced, tiny, irregular shrinkage cracks, which are approximately at right angles to the grooves. Many cracks are less than 1 mm. in depth, and some have an edge serrated by minute teeth. The cavern in which the specimen occurs is within a few feet of the original surface of a flow of basalt not less than 90 feet in depth, which is well exposed in a quarry close at hand; it [illegible] represent a "gas blister."

At Newmarket, grooves almost identical with those just described appear on a surface approximately 4 by 3 feet in size, which is part of a thrust plane, dipping 30°. It forms a portion of the roof of a cavern about 5 feet across and 7 feet high, which is located 10 feet below the present surface of a flow of basalt not less than 50 feet in depth, which has come from Mount Eden volcano. This cavern, like that at Auckland Grammar School, probably is a gas blister in the lava.

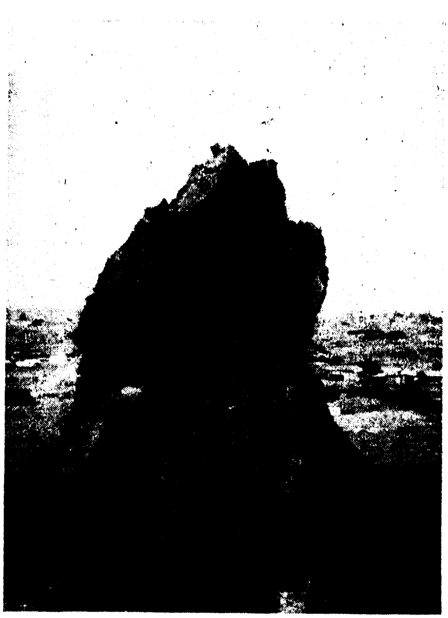


FIG. 1.—Basalt dike locally with a median cavity, Mount Wellington, Auckland. Main dike on left; a branch on right. The hat rests on scoriae and serves to give the scale.

The origin of the grooves in the two last specimens is clear from the demonstration of thrust in the occurrence at Newmarket and is similar to that of grooves from McCartys flow, New Mexico, which Nichols (1938) has shown to have been formed by the upthrust from below of semisolid, pasty lava upon the rough surface of upper, wholly solid, cooled rock. It is clear from the details of the surface of the Auckland Gram-

mar School specimen that the upthrust mass must have reached the stage of almost complete solidity when movement occurred; the transverse cracks otherwise would have deepened as cooling progressed.

ACKNOWLEDGMENT.—The writers are indebted to Mr. H. R. Kennedy for the grooved lava from Mount Wellington and also for drawing their attention to interesting features of the quarry at that place.

## REFERENCES CITED

BARTRUM, J. A. (1928) Lava slickensides at Auckland: New Zealand Jour. Science and Technology, vol. 10, pp. 23–25.

NICHOLS, R. L. (1938) Grooved lava: Jour. Geology, vol. 46, pp. 601–614.

——— and STEARNS, C. E. (1940) Grooved lava in the cross section of Big Craters, Idaho: Am. Jour. Sci., vol. 238, pp. 22–31.

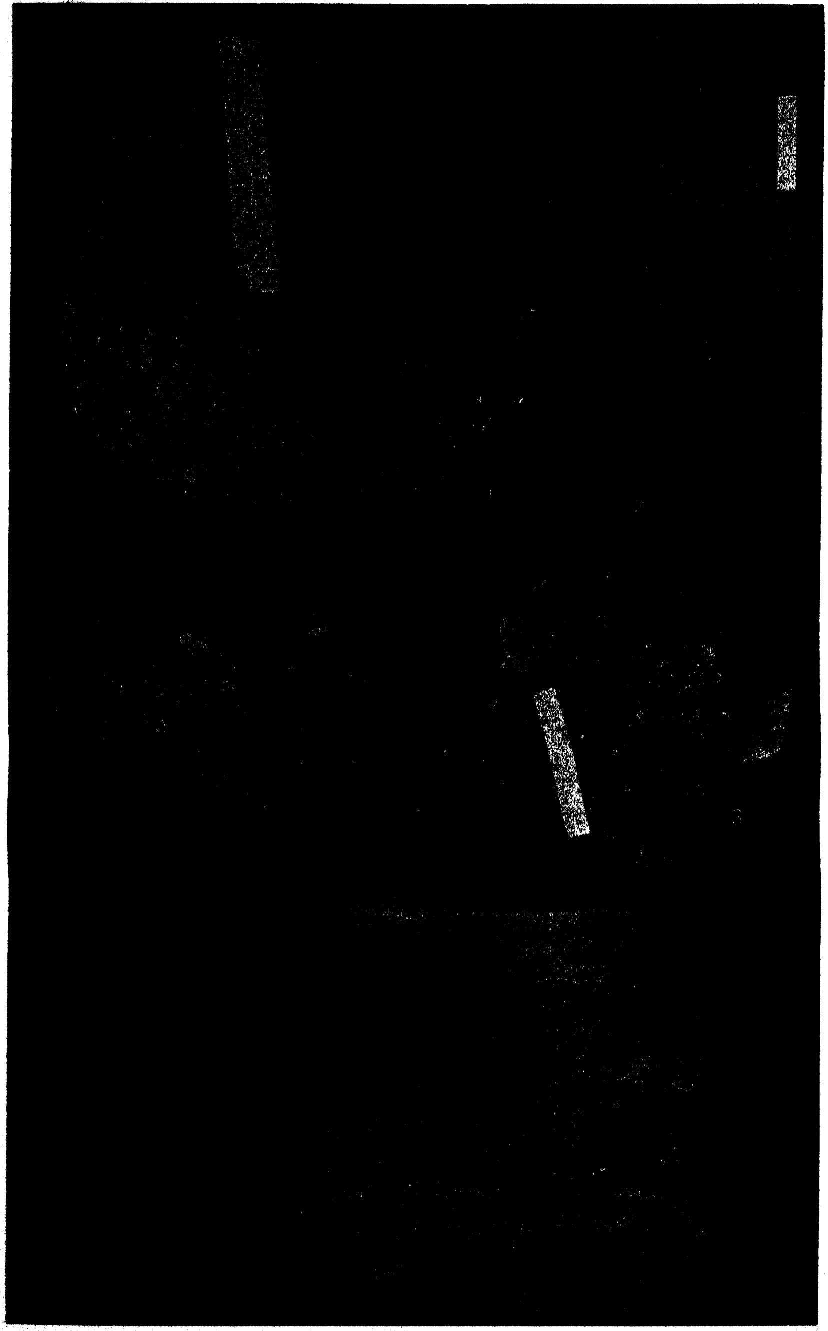
## PLATE 1

FIG. 1.—End-view of the spiracle tube of fig. 3, showing the fused lining of the spiracle itself.

FIG. 2.—Re-fused glazed surface of wall of cavity of dike of fig. 1. Drip features are clearly visible.

FIG. 3.—Spiracle-tube of agglutinated scoriae, Mount Wellington, Auckland. An opening of the spiracle shows on the right.

FIG. 4.—Grooved lava, Mount Wellington, Auckland.



# DISCUSSION

## MIMA MOUNDS

CHAPMAN GRANT
San Diego, California

Dr. Victor B. Scheffer published an interesting article (1947) on the origin of the "mounds" which are to be found throughout much of the western plains. The title, *The Mystery of the Mima Mounds*, leaves us to guess whether or not he solved the riddle. In his text Dr. Scheffer is satisfied that they were built by pocket gophers. In an earlier article, which he wrote with W. W. Dalquest (1942), the case is more convincingly stated. His arguments may be paraphrased somewhat like this:

In deep soil the pocket gopher digs his tunnels, lines his underground nest, and lives his life without causing any changes in the landscape. If, however, there is a hard substratum within a foot or two of the surface, his actions are entirely different. A nest site is then dug deep into the hard stratum. This done, the gopher digs radiating tunnels for foraging. He brings a minute excess of soil *toward* the nest site over what he takes *from* it. After thousands of years his descendants have accumulated a "mound" at the original nest site. His progeny has peopled the entire shallow-soiled prairie as thickly as the forage will permit, and this proximity governs the spacing of the mounds as they are now found. Cobbles which occur in the hill are gradually settled by his digging under them to construct a deep nest for his young—safe from wandering bear, wolf, and wildcat! In excavating his foraging tunnels between mounds, he is motivated by a less vital instinct, so, having no reason to go deep to protect his young, he merely passes around any cobble encountered, and, because the soil is gradually moved to the hill, cobbles are exposed and strew the intermound spaces. All mounds are gopher-made whether gophers now inhabit the area or not. Shallow groundwater may have the same effect as a hard substratum.

The following points made by Dr. Scheffer (1947) are followed by the writer's criticism.

1. At certain places [the gophers] dug deeply into the gravelly subsoil in order to make nest chambers.

Years of experience with the pocket gopher has convinced me that its nest chamber is no deeper than most of its burrow. Dr. Scheffer's figure 2 illustrates the shallowness of the nest.

2. Because of depth, the nest chambers are ". . . well protected from prowling bear, wolf or wildcat."

The chief protection of the gopher, like that of the rabbit, is its fecundity. In addition, it tries to stay underground and keep its burrow well plugged. Dr. Scheffer apparently thinks that predators seek the young in preference to adults. The number of nests destroyed by bear or wolf must be very small compared to the adults destroyed by these two animals. The number of nests dug out by wildcats is 0. A wildcat does not dig. The real vertebrate enemies of the gopher are snakes, owls, hawks, and weasels. Disease and parasites, drowning and starvation, are the effective checks on this rodent—not bears.

3. "Areal spacing of the nest chambers corresponded to the size of the 'territory' of each animal. The center of an old territory now marks, we believe, the center of a modern mound."

The territory of a pocket gopher does not normally radiate from a center but is, instead, a rather long, narrow figure, or a line. Nowhere are territories found as close together as are the mounds, or as evenly spaced. The evenness of distribution is too great in that it would require a vast pasture, rich enough to support an evenly concentrated population. In reality, any such expanse would contain areas of poor pasture, where mounds would have been spaced farther apart or would be missing.

4. "When the animal ran into a large boulder it undermined the obstruction and allowed it to settle."

This assigns a complicated purposeful action on the part of the gopher in excavating its nest

site. There is no evidence to support this statement. The use of the term "boulder" in many places in Dr. Scheffer's article is incorrect. The author means "cobblestones."

5. "Thus, we now find, at the base of most mounds, a concentration of coarser materials."

Dr. Scheffer tries to prove a point uncalled for by his own theory. His gophers are supposed to have started upon a flat plain and to have ultimately piled up mounds containing pebbles "no larger than walnuts." He now attempts to prove how the gophers got "boulders" *out* of mounds which could not have contained anything larger than a walnut in the first place!

6. According to Dr. Scheffer, when the gopher dug tunnels for foraging, it was driven by less powerful instincts than that of nestbuilding, so, when it encountered a rock, it simply passed around it. I do not believe that a gopher acts differently toward an obstruction in a runway or a nest site.

7. ". . . shoving dirt along as it went."

Apparently the object of this statement is to prove that gophers might push more dirt in one direction than in another. Actually, gophers make "spoil" dumps at more or less regular intervals along their rather straight line of burrowing. There is no evidence that the dumps are placed in any position or direction other than that dictated by convenience.

8. "Thus, we find plainly exposed in the intermound hollows large boulders that were doubtless at one time buried in the topsoil."

This would be possible only if the gophers continually made their spoil dumps toward a center. There is no reason to believe that such was the case. They certainly do not do so at this time.

9. ". . . 'mound roots' . . . are simply abandoned gopher tunnels now filled with black silt. . . . They call to mind the peculiar devil's corkscrews . . . [of] Nebraska . . . [which] are now generally believed to be the casts of burrows of extinct rodents."

In a footnote Dr. Scheffer refers to A. L. Lugn (1941, p. 673). Lugn, however, stated: ". . . vegetal origin is believed demonstrated . . . [as the cause of the corkscrews]."

10. "Where [a] nesting chamber collapsed [it] caused a depression at the crest of the mound, a characteristic feature of many of the mounds. . . ."

Dr. Scheffer illustrates everything else, even including a Fresno scraper, which is used to grade down the mounds for agricultural purposes, but omits illustrating this special depression. I have never seen such a depression and would have appreciated a picture of one. A nest chamber is the size of the bowl of a derby hat or smaller. It is never reused or enlarged, to my knowledge. Gophers do not seek out high ground for nest sites.

11. "In fancy, it is easy to picture the start of a Mima Mound."

Without proof or reason to believe that gophers move an excess of soil toward a common center, it is impossible to fancy the beginning of a mound.

12. "It is less easy to account for its [the mound's] growth."

I see no difference between the inception and the later growth of a mound. Once started, I can visualize its continued growth. Why the gophers should stop building at a certain size might be a problem; but then, muskrat nests are of approximately the same size.

13. "For reasons that may never be known, the gophers carried more dirt towards the nest than away from it."

I agree with the first part of this sentence but not the last part. In fact, I am speechless at this admission and wonder why the article was written.

Dr. Scheffer's article ends with some "conclusions" which I treat similarly to the body of the article:

14. ". . . mounds are distributed . . . exclusively in the range of the pocket gopher."

According to his own statement: "There are no gophers on the Mima Prairie."

15. "Burrowing animals with habits similar to those of the gopher, namely the ground squirrel (*Citellus*) and the mole (*Scapanus*) . . . are not pertinent to the formation of mounds. . . ."

According to his first conclusion, there are no gophers in the Mima Prairie, so they would not be pertinent either. Dr. Scheffer seems to propound a theory that the range of a species is

rather permanent. Animals as well as plants migrate with changing conditions.

I am amazed at the statement that the burrowing habits of the squirrel, mole, and pocket gopher are "similar." I cannot name three better examples of diametrically different uses of the earth by digging animals. The ground squirrel lives in colonies, burrows only for protection, and forages above ground by day. It has good eyesight, moderately developed nails for digging, and does not use its teeth for digging. The spoil is deposited at the mouth of the unplugged burrow. The mole, nearly blind, possesses a remarkably specialized body for forcing its way along under the sod in search of worms and insects. It seldom burrows, does not leave spoil dumps along its unplugged tunnel, and does not forage above ground. The pocket gopher plugs his burrow and can feed only over a radius of his body length around a forage hole—never completely emerging. (The young and males do travel above ground at certain seasons at night to start a new territory or to find a mate.) Its teeth and toenails are modified for digging and its pouches for carrying soil.

16. ". . . mounds are found only where . . .a thin layer of workable soil [overlies] a dense substratum."

Dr. Scheffer quotes Vernon Bailey as stating that mounds occur in southwestern Louisiana. There is no hard substratum there.

17. ". . . in deep sandy soil . . . [gophers] never form Mima-type mounds."

There is no proof that they produce mounds in shallow soil either. There are great areas of mounds where no gophers occur and vice versa.

18. According to Dr. Scheffer, the mounds are not deposits since they are unoriented and occasionally occur on slopes.

Dr. Scheffer's figure (1947, p. 286) shows that the mounds have a marked orientation. The fact that the mounds differ in texture from their bases proves that they were *built* by some means.

19. According to Dr. Scheffer, the mounds are not due to erosion because the interspaces are frequently closed depressions.

The most obvious disproof of dissectional residue is that the substratum differs from the mounds.

## REFERENCES CITED


DALQUEST, W. W., and SCHEFFER, V. B. (1942) The origin of the Mima Mounds of western Washington: Jour. Geology, vol. 50, pp. 68–84.

LUGN, A. L. (1941) The origin of Daemonelix: Jour. Geology, vol. 49, pp. 673–696.

SCHEFFER, V. B. (1947) The mystery of the Mima Mounds: Sci. Monthly, vol. 65, pp. 283–294.


# MIMA MOUNDS: A REPLY


VICTOR B. SCHEFFER

U.S. Fish and Wildlife Service, Seattle, Washington


Major Grant believes that gophers behave in one way, and we[1] believe that they behave in another—or we admit that we do not know exactly how they behave. I have plainly stated (1947, pp. 293, 294) that our evidence is indirect; that we have not seen gophers building a giant mound; that we do not know whether mound building is a contemporary or a historic process; and that we do not know whether the stimulus for mound building is a hardpan or a high water table or both.

Our main contentions are (1) that mounds of the Mima type occur only within the range of

[1] In 1942 Walter W. Dalquest and I developed the theory of origin of the Mima Mounds by gopher activity. Since Mr. Dalquest is in Mexico and unable to enter the present discussion, I am taking the liberty of defending his views as well as my own.

gophers, living or extirpated, and (2) that only a living, adaptable force, not a physical agency, could have produced the Mima-type mounds out of widely varying materials and in widely varying environments from Mexico to Puget Sound. We note that Chapman Grant does not propose an alternative theory for the origin of the mounds.

In his introduction Major Grant has paraphrased our arguments quite well. We disclaim the statement: "All mounds are gopher-made whether gophers now inhabit the area or not." There are, of course, many kinds of mounds. We claim only that the Mima-type mound or pimple mound, as illustrated in our article, is gopher-made. Furthermore, the only place on the West Coast where we have seen Mima-type mounds unassociated with living gophers is Mima Prairie, a small opening of perhaps 10 square miles. Major Grant, later in his paper, emphasizes the fact that there are no gophers here. As we have explained (1942, p. 81; 1947, p. 293), the absence of living gophers on this specific prairie is unimportant. Rather than selecting Mima Prairie as the type locality of the pimple mounds, we could as easily have selected Tenino Prairie, where there *are* mounds inhabited by living gophers, 1 mile southeast of Mima Prairie.

We shall attempt to answer Major Grant's criticisms, numbered for easy reference, as follows:

1. The depth of a gopher nest varies. The nests with which Major Grant is familiar may be shallow, but

> . . . on the gravelly prairies of western Washington the feeding runs of [the gopher], as they approach the vicinity of the nest, descend almost vertically to depths of 2, 3, and even 5 feet. . . . In excavating four burrow systems of this species, the writer found the nests at depths of 26, 29, 34, and 36 inches, respectively [Scheffer, 1931, p. 13].

2. The specific enemies or adverse conditions that a gopher avoids in nest building are unimportant in our discussion. The pertinent facts are that a gopher always builds a nest, the nest is the focal point of the home territory, and the nest is deeper than the average foraging runway. Fish and Wildlife Service records of stomach examinations show, however, that gophers *are* eaten by bear, wolf, and wildcat; but, again, these facts are unimportant.

3. Major Grant raises an interesting question in connection with the shape and spatial relations of the foraging territory. It is true that the foraging territory, or burrow system, is more linear than circular, *at any given time*. Enough burrow systems have been excavated to prove this point. The burrow system is constantly changing, however, as the gopher searches for plant roots. New tunnels are made, old ones are filled; a gopher dies and another takes its place; subadult gophers leave the nest and seek new territory. We believe that the effect over many gopher generations is a honeycomb-like spacing of the mounds.

In this connection the areal distribution of other mammals is significant. On the Pribilof Islands, Alaska, the male fur seals gather in the spring, each taking up a territory on the beach and jealously guarding it from newcomers. The seals are not spaced with the regularity of checkers on a board, and yet they are certainly not spaced at random. Here we have a visible example of the fairly uniform spacing of family territories. We cannot see the spacing of gophers because they live underground, but we can infer that the individual territory tends to be circular or, more precisely, hexagonal in shape. The tightness or looseness of the network of home territories probably varies with the kind of soil and vegetative cover; that is, a gopher family requires more foraging ground where food plants are scarce than where they are plentiful.

In previous accounts we have stated our belief that the mound is developed around the center of an old nest burrow. We do not mean to imply that each mound is still the hereditary castle of a family of gophers. At any given time some mounds are occupied, and others are not. Were all the mounds occupied at once, Major Grant could reasonably feel that the gophers were overcrowded.

4. We did not state or imply that the gopher uses reasoning power. When it moves dirt from the side or top of a boulder, the boulder remains at rest. When it moves dirt from beneath the boulder, the boulder tends to settle. The evidence is a layer of coarser materials at the base of the Mima-type mound.

5. The intermound cobbles, or boulders, on the Puget Sound prairies were not moved out of mounds by gophers. They are more or less *in situ*, although many of them have been bared by the removal of silt gravel.

6. We feel that this is a matter of opinion.

7. Major Grant has understood us correctly. We believe that, where soil and climatic conditions are favorable to mound building, gophers *do* push more dirt toward the nest. In addition to mounds, of course, one can see the small spoil

heaps or gopher hills scattered on and between the mounds.

8. It is true that gophers do not everywhere make their spoil dumps toward the center. We believe, though, that over periods of time, in the shuttling of dirt as the gopher digs for plant roots and for nest and food-storage chambers, there is a differential movement of materials favoring the growth of the mound.

9. We inadvertently cited Lugn. We should have cited C. Bertrand Schultz who, in 1942, stated that "most palaeontologists now believe . . . that Daimonelices are the casts of rodent burrows."

10. A shallow depression occurs on the top or flank of many mounds on Mima Prairie. On this prairie, the reader will recall, there are no living gophers. We interpret the depression as the collapse of an old nesting chamber somewhere in the mound. It may, however, represent the recent activity of moles or livestock. J Harlen Bretz (1913, p. 101) referred to "occasional . . . sunken areas a foot or so across on these mounds. . . . The small sunken areas are so recent that the sod has not healed over the marginal cracks."

Major Grant states that "gophers do not seek out high ground for nest sites." We have been given two photographs taken in Texas, showing mounds sliced open to reveal nest chambers well above the surrounding ground level. Here winter flooding evidently obliges the animals to build their nests out of danger. Pennoyer F. English (1932, p. 127, pl. 9) has published a photograph of a similar Texas mound, with the statement that the gopher here "builds its nest not deep in the ground but in an enormous mound."

11. We find it easy to fancy the beginning of a mound as a center of activity in the vicinity of the nest. Others, with equal freedom, may fancy the beginning of a mound as a platform on which the gopher attempts to raise its nest out of the mud at a certain season.

12. The size of a Mima-type mound in a particular locality probably depends upon many factors. We are more concerned, though, with establishing the fact that the mounds are of gopher origin than with the ultimate size to which the mounds may grow. The industry of the gopher as a mover of soils is perhaps greater than many realize. According to Lincoln Ellison (1946, p. 113):

> In what is considered to be a representative part of the subalpine zone of the Wasatch Plateau in central Utah, annual displacement of soil to the surface by pocket gophers was found in 1941 to be at least 5 tons per acre and to cover 3½ per cent of the surface. The base population of pocket gophers is estimated to be somewhere between 4 and 16 animals per acre.

13. See No. 3.

14. See our introduction, paragraph 3.

15. We do not "propound a theory that the range of a species is rather permanent." In fact, we once published a paper describing the migrational history of gophers (1944, pp. 308–333, 423–450).

Use of the word "similar" here is a semantic privilege. The mole, ground squirrel, and pocket gopher are similar in that they make tunnels and throw out excavated soil in mounds.

16. We have clearly posed the question (1947, p. 294): "Does ground water at certain times of the year and in certain localities act in the same way that a soil hardpan does. . . ?" We have not seen the mounds of Louisiana, but, from the evidence of aerial photographs and from correspondence with Vernon Bailey, we believe them to be of the Mima type. *If* they actually are of the Mima type and *if* there is no hard substrate there, the water table may act as hardpan does on certain other mound prairies.

17. Major Grant states that "there are great areas of mounds where no gophers occur and vice versa." We have pointed out that the absence of living gophers in Mima-type mounds does not invalidate the theory of their origin in past years by gophers. We have also pointed out that there are many smooth prairies where gophers are living but where conditions do not favor the formation of mounds (1942, pp. 81, 84; 1947, p. 293).

18. Bretz's map, which we used as a figure (1947, p. 286), perhaps suggests slight orientation. Bretz himself stated that, while there is commonly an elongation of the mound, it "does not conform to any definite orientation" (1913, p. 83).

Major Grant discusses orientation, but he does not comment on our point that the mounds occasionally occur on slopes. Here is a phenomenon difficult to account for, unless one accepts the theory that the mounds are built from within, by animals. We quite agree with his statement: "The fact that mounds differ in texture from their bases proves that they were *built* by some means."

19. Major Grant's criticism is not clear.

ACKNOWLEDGMENT.—Dr. J. Hoover Mackin has read the manuscript and offered helpful suggestions.

## REFERENCES CITED

BRETZ, J HARLEN (1913) Glaciation of the Puget Sound region: Washington Geol. Survey Bull. 8.

DALQUEST, W. W., and SCHEFFER, V. B. (1942) The origin of the Mima Mounds of western Washington: Jour. Geology, vol. 50, pp. 68–84.

——— — ——— (1944) Distribution and variation in pocket gophers, *Thomomys talpoides*, in the state of Washington: Am. Naturalist, vol. 78, pp. 308–333, 423–450.

ELLISON, LINCOLN (1946) The pocket gopher in relation to soil erosion on mountain ranges: Ecology, vol. 27, pp. 101–114.

ENGLISH, P. F. (1932) Some habits of the pocket gopher, *Geomys breviceps breviceps:* Jour. Mammalogy, vol. 13, pp. 126–132.

SCHEFFER, T. H. (1931) Habits and economic status of the pocket gophers: U.S. Dept. Agr. Tech. Bull. 224.

SCHEFFER, V. B. (1947) The mystery of the Mima Mounds: Sci. Monthly, vol. 65, pp. 283–294.

SCHULTZ, C. B. (1942) A review of the Daimonelix problem: Univ. Nebraska Studies in Science and Technology 2.

# PETROLOGICAL NOTES AND REVIEWS

## RECENT CONTRIBUTIONS TO THE GRANITE PROBLEM

TOM. F. W. BARTH
University of Chicago

During the last one hundred and fifty years the question of the genesis of granite has frequently reappeared as the chief subject of discussion among geologists and geophysicists. Among the recent contributions are those of R. Perrin and M. Roubault, P. Niggli, M. Reinhard, E. Raguin, and H. G. Backlund.

Two articles published in English (Grout, 1941; Read, 1943, 1944) are presumably easily available and are not reviewed here.

*Om granit och gnejs och jordens ålder* ('On Granite and Gneiss and the Age of the Earth'). By H. G. BACKLUND. ("Kungl. Vetenskapssocietetens. Arsbok, 1947.") Uppsala, 1947. Pp. 67.

Backlund's book, one of the most recent, is reviewed first because Backlund, probably more than anyone else, was responsible for reviving the discussion of the granite problem in 1936 (pp. 293-347) by propounding a granitization theory disclaiming the primary, magmatic character of the rapakivi granites and similar rocks. For more recent references see Backlund's papers of 1938 and 1946.

It is a puzzling fact that the mineral assemblage of granite cannot be duplicated in laboratory furnaces, although the frequency and large areal distribution of granites in orogenic areas of all ages indicate that they form with great ease in nature.

It is of genetic importance that most ore deposits, oxidic as well as sulphidic, are intimately associated with granites in the upper parts of the earth's crust. But the exact relationship is obscure—often the practical prospector has more success than the professional geologist in discovering new ore (Rastall, 1945).

Pegmatites are closely related to granites. Less than fifty years ago, a leading scientist, A. E. Nordenskiöld, said that no geologist in his right mind could believe in the formation of pegmatites from molten solutions. Subsequently, they were explained as late crystallization products of granite magma. Now great quantities of volatiles, especially water, are believed to be instrumental in their formation (pneumatolysis). Some pegmatites carry rare elements. In this respect they are analogous to ore deposits, especially where large parts of them consist of only one "rare" mineral (petalite, amblygonite, pollucite, beryl, or others). The distribution of such minerals is very irregular, and many pegmatites are quite "sterile."

Migration of elements from the granite across the boundary into the adjacent rocks has taken place, contrary to the opinions of Rosenbusch and his school. The granite is not a passive carrier of heat; it is an active vehicle which conveys heat and material, sets up temperature gradients, and mobilizes gases and vapors and other chemical ions, thus inducing minerochemical changes in vast areas far outside the contact face (metasomatism). The changes take place volume by volume. Homogeneous mineral assemblages (often monomineralic) may form, which are lower in silica than either the granite or the adjacent rock. They represent geochemical "culminations," are called "diabrochites" (Backlund, 1943; Reynolds, 1947), and have often been misinterpreted as truly magmatic.

According to the "reaction principle" developed in the laboratory, the early products of crystallization cannot become assimilated by the rest magma; but the gradual disappearance of aggregates of early crystals and of basic inclusions in all granite areas is well known. The cations Fe and Mg show a tendency to migrate. The evolution of the series: greenschist → amphibolite → metabasite, which is closely connected with the gneiss-granite problem, cannot be explained through the action of hot magmas from great depths.

The "ichor" of Sederholm is redefined and

called "emanation" (Wegmann, 1935). If a rock, $A'$, is granitized into $B'$, the reaction in its first approximation may be written as follows:

$$A + x = B ,$$

in which $A$ and $B$ are the specific gravities of educt and product, respectively, and $x$ represents the quantity of the emanation. The relation of $x$ is as follows:

$$x = e_1 - e_2 - a_1 ,$$

in which $e_1$ is introduced emanation, $e_2$ is the fraction of $e_1$ not fixed in $A'$, and $a_1$ is the fraction given off by $A'$ in the reaction $A' \rightarrow B'$. The emanation is of pulsating character and represents a migration of ions facilitated by lattice disorders, substitutions, polymorphic inversions, lattice deformations, etc. The process is very slow even in comparison with exogenic processes, such as denudation, erosion, etc.

Backlund is a staunch advocate of the uniformitarian view and claims that all the extensive and still largely undifferentiated pre-Cambrian gneiss areas represent relics of old mountain chains. (There have been between twelve and fifteen different orogenies in the history of the earth.) During each orogeny, intense granitization took place. Large parts of mountain chains that originally were made up of normal sediments were replaced by granite and gneiss, regardless of age, in all geological eras. This was not a replacement of solid rock by molten magma or a melting of the sediments by magmas, for the amounts of missing sediments are much too large, the volume relations too acute, and the replacing rocks too homogeneous. The various radioactive age determinations (made on granitic or pegmatitic minerals) simply indicate the time of the several orogenies. But there are always sediments older than the orogeny, even biogenic sediments. Thus the total age of the earth cannot be determined in this way.

Sediments of granitic composition (arkoses) are most easily granitized (corresponding to small values of $x$ and $a_1$). If the chemical differences are large, then $a_1$ will be large, and the original rock will long resist granitization. This corresponds to the following (relic) series in granitization (in order of increasing ability to survive): (1) mica schist, (2) quartzite, (3) limestone, (4) basites, (5) ores.

Of particular interest is the distribution of the trace elements, which are said to harmonize with a diffusion differentiation but not with a magmatic differentiation (Bray, 1942; Sahama, 1945*a*, 1945*b*; Sandell, 1943; Shimer, 1943).

The continuous succession of granitizing processes may be presented as shown below. Backlund praises Sederholm, who, although he was a pupil of Brögger and Rosenbusch, undertook the further development of the granitization theory after the French had become silent at the beginning of the present century. Sederholm's grand role, in the face of united opposition for several decades, cannot be overrated. Today Sederholm's ideas are pushed further than he ever had supposed possible.

*Le Granite et les réactions à l'état solide*. By R. PERRIN and M. ROUBAULT. (Bull. du Service Carte Géol. de l'Algérie 5e sér. [Petrographie] No. 4.) 1939. Pp. 182; figs. 8; pls. 6.

The authors are worthy successors of the splendid French school of 1847 and of the skilful petrographers of the eighties and nineties: Michel-Lévy, Termier, Barrois, Lacroix, creators of the striking *tache d'huile* and *colonnes filtrantes* analogies for describing the granitization processes. In 1937 they published a book giving the underlying physicochemical theory of the reaction and the diffusion processes in the solid state and the general geological conclusions derived from them. The authors are convinced that the existence of selective diffusion is the clue to all problems of metamorphism. The origin of granite is explained in this way.

Rocks form by reactions between solids or by crystallization of a magma. The problem is to find out which process has been active in the formation of any given rock.

GRANITIZING PROCESSES

$$\begin{tikzcd}
 & & & \begin{array}{c}\textbf{extrusion}\\ \textbf{(acid lava)}\end{array} \\
 & & \begin{array}{c}\textbf{rheomorphism}\\ \textbf{(granite II)}\end{array} \arrow[r] \arrow[ur] & \begin{array}{c}\textbf{intrusion}\\ \textbf{(granite I)}\end{array} \\
\begin{array}{c}\textbf{sedimentation}\\ \textbf{(sediment)}\end{array} \arrow[r] & \begin{array}{c}\textbf{metamorphism}\\ \textbf{(metamorphite)}\end{array} \arrow[r] & \begin{array}{c}\textbf{metasomatism}\\ \textbf{(gneiss)}\end{array} \arrow[r] \arrow[u] & \begin{array}{c}\textbf{granitization}\\ \textbf{(granite I)}\end{array}
\end{tikzcd}$$

By thoroughly searching the geological literature, the authors have found the following statements in favor of a magmatic origin of granite. They are all rejected, however.

1. *Order and temperature of crystallization.*—The authors think that irregularities in the order of the elements and certain phenomena of zoning are incompatible with crystallization from a melt.

2. *Injection "lit par lit."*—The hypothesis, in accordance with Bowen's reaction principle, of a residual liquid which is mechanically injected into schist layers without dissolving the basic minerals is, according to these authors, contrary to observations. They conclude, therefore, with Michel-Lévy, that the phenomenon is metamorphic in origin.

3. *The phenomena of inhibition and soaking.*—The development of granitic minerals in enclosed blocks is best explained by reaction in the solid state.

4. *The existence of dikes and apophyses.*—This is given a long and interesting discussion with numerous references to the observations and interpretations of many leading geologists. Despite the subjective impressions of earlier observers, the authors conclude that the dikes and apophyses are not proof of a liquid granite. They merit renewed and precise study, however, and individual consideration of each case.

5. *The transition of granite into microgranite and rhyolite.*—Are the changes in texture due to a change of the rate of cooling of the magma or to subsequent recrystallization? The authors favor the latter hypothesis.

6. *Homogeneity of granitic massifs.*—In the opinion of these authors, homogeneity does not support the differentiation theory of Bowen. It should be added that the homogeneity is only relative: analyses from various parts of a massif are usually similar but not identical.

7. *Presence of liquid inclusions.*—Liquid inclusions and granites formed in the solid state are not incompatible.

Among the additional problems which are thoroughly discussed are the digestion of dikes, the gradual transition of schist or gneiss into granite, the *mise en place* of the granite, reactions in the solid state in massif rocks other than granite, and the possibility of large-scale solid diffusion.

In summing up, the authors believe that the evidences in favor of a liquid granite in all cases rest on "subjective impressions" and on analogies associated with the molten state. But detailed studies prove that reaction in the solid state can make, and has made, similar rock bodies. The authors were unable to discover a single clear case in which a granite showed a magmatic mode of origin. On the other hand, numerous granites, varying greatly in geology and in mode of occurrence, show evidence of formation in the solid state.

The book is a store of information; it is stimulating and full of new ideas. It is logically and clearly written, and there are no wild speculations. Each statement is carefully considered and supported by thoroughly checked physicochemical facts. It is a pioneer work of great value to geological thought.

"Das Problem der Granitbildung." By P. NIGGLI. (*Schweizer. min. pet. Mitt.*, vol. 22). 1942. Pp. 84.

It is almost impossible to give a review of Niggli's paper. It is in itself a review, although of great length. With the help of very rich historical material, the development of the various notions pertaining to the genesis of granites is discussed.

The modern controversial subjects have received fair treatment, in that long sections (several pages and more) of a great number of contemporary writers are quoted verbatim in the original language.

But there can be no mistake about the point of view taken by Niggli himself. His paper is a clear vindication of the purely magmatic origin of granite. He has no use for concepts like granitization, "intergranular films," or "migmas"; even the term "migmatite" as a rock name is grudgingly accepted. He points out that the views put forth in his paper are very similar to those expressed by Grout (1941), whom he quotes at length and with approval.

"Die leukogranitischen, trondhjemitischen und leukosyenitischen Magmen und die Anatexis." By P. NIGGLI. (*Schweizer. min. pet. Mitt.*, vol. 26.) 1946. Pp. 34.

The attitude adopted in this later article is that each granite should be considered on its own merit, and its particular origin discussed. In the origin of granite there is no "either-or." Niggli is now willing to accept the existence of at least three kinds of granite: "magmagranit," directly crystallized from a magma; "metagranit" (as used by Kranck in Wegmann and Kranck, 1931), formed by simple recrystallization of arkose and similar rocks; and "migmatit-

granit" (as used by Kranck in Wegmann and Kranck, 1931), representing the end-product of the metasomatic series: sediment—metamorphite—ultrametamorphite—migmatite granite.

The following definitions, which are partly adopted from Rittman (1942–1945), are proposed:

*Migma* is a silicate mass, generated through partial deliquescence of rocks, containing relic minerals (palaeosoms), and unable to be intruded like a magma.

*Syntectic magma* has changed its composition through assimilation.

*Palingenic magma* is of migmatic origin.

*Hybrid magma* is the result of the mixing of two or more magmas.

*Parautochthonous magma* generates and congeals more or less in the same place.

*Allochthonous magma* generates at some place other than that where it congeals, e.g., at great depth.

Niggli's superior insight and intimate knowledge of the subject make his reviews sources of information which should be consulted by all students of the controversial granite-gneiss problem.

*Über die Entstehung des Granits.* By M. REINHARD. ("Basler Universitätsreden" 16.) Basel: Helbing & Lichtenhahn, 1943. Pp. 38.

The fundamental queries of the magmatic petrogenesis are: How did the individual rocks and rock series develop, and why is granite absolutely dominant among the deep-seated rocks, while basalt dominates among the extrusive rocks?

There are three different ways to form a granite:

1. A residual granitic liquid will develop through fractional crystallization of a basaltic magma.

2. The granitic rest magma, rich in volatile constituents, is eventually replaced and succeeded by a fluid phase high in alkalies and silica. This phase will react with the minerals of the congealed granite and partly change them into new minerals. It will also diffuse into the adjacent rock and react with its minerals. Sandstones and shales may thus be changed into rocks, which in their mineral composition and structure are identical to granite.

3. The deeper parts of all orogenic zones are dominantly granitic in composition (massive granite, schistose gneiss, augen gneiss, layered gneiss, injection gneiss, migmatites, every transition from truly magmatic rocks to metamorphic sediments). This is the great melting-pot. The process of melting or solution is the opposite of fractional crystallization. The material that crystallizes last will go into solution first. Not wholesale melting, but a differential anatexis, takes place, with formation of a granitic ichor, or migma, which subsequently soaks and permeates the whole rock complex (the name "migma" was coined by Reinhard in 1934).

It is usually impossible to determine the mode of origin of a granite body because the physical and chemical conditions deep in the lithosphere are very inadequately known.

Juvenile or palingene? Magmatic or metamorphic? To the first query we have no answer —it is a question of belief, not of science. We should not contaminate our science with useless discussions but should show tolerance. To the second query we may find a solution through careful studies in the field and in the laboratories. But the question is not rightly presented. It is not "either-or." Most granites are of mixed origin, and the problem is to establish the extent to which they are magmatic or metamorphic. For such studies our conservative petrographic classification is not wide enough. The author favors a fourfold classification: igneous, migmatic, metamorphic, and sedimentary (Barth, in Barth, Correns, and Eskola, 1939).

Studies for twenty-five years in Wallis and Tessin have demonstrated that rocks which were first regarded as orthogneisses and granite gneisses actually are the results of metasomatic processes. The objection of Bowen, that it is easy to invoke "gaseous transfer" because it may explain everything, and of Niggli, that a general gassing and emanation theory is uncontrollable, do not constitute arguments against a prevalence of metasomatic processes.

Reinhard's book gives the full historical background of all these ideas and is a dignified, objective, and logical contribution to one of the most fundamental problems in contemporary geology.

*Géologie du granite.* By E. RAGUIN. Paris: Masson & Cie, 1946. Pp. 211; figs. 46.

Because of the war the book by Perrin and Roubault did not get a good distribution outside France. But in France it was known and

naturally met with some opposition. Raguin's book testifies to this.

Raguin thinks that the geology of granite, which is defined as an "eruptive" rock, is of fundamental importance for both ore geology and orogenesis and consequently for the study of the evolution of the deeper zones of the globe. The author draws his knowledge from many sources, and his bibliography includes 126 references from many countries.

However, the physicochemical point of view is put aside; there are no chemical analyses, diagrams of differentiation, or discussion of laboratory experiments. The geological viewpoint prevails, that is, the viewpoint of the "naturalist" who is interested in the crust of the earth.

The book starts with a clear, concise description of the minerals of granite, order of crystallization, structure and texture, and classification of granites, followed by a discussion of the diversity of granite, which suggests a diversity of origin.

There are two types of granite: anatectic and intrusive, bridged by transitional types. Anatectic granites may grade into gneisses and migmatites; intrusive granites form batholiths or other large massifs.

The structural behavior of granitic bodies is adequately treated, including schistosity, foliation, cleavage, jointing systems, and general tectonics. Alteration of granites, crushing, and mylonitization are also amply discussed.

In the formation of granite, great importance is placed on assimilation by the magma. The differentiation of granitic magma is also discussed, but the author fails to explain the relatively great abundance of granite magma compared with any other type of intrusive magma. Because granitic lavas (rhyolites, etc.) probably make up less than 1 per cent of all visible extrusive rocks, the question must be answered.

Dike rocks associated with granite (aplites, pegmatites, pneumatolytic veins, lamprophyres, and porphyries) are described.

The chapter on the metamorphism of granite is very interesting. The underlying ideas are those of the great French school of metamorphism (Termier, Michel-Lévy, Barrois, etc.), supplemented by the work of the Fenno-Scandian school (Sederholm, Holmquist) and the Swiss geologist, Wegmann. From his wide readings the author also incorporates important conceptions from Becke, Harker, Niggli, and Sander, thus arriving at a polished and consistent picture of the metamorphism. However, he rejects the possibilities of reaction and diffusion of chemical substances in the solid state, ideas that have been recently and independently presented in France by Perrin and Roubault, in Great Britain by H. H. Read and Doris Reynolds, in Scandinavia by Backlund and by the petrographic school at Oslo.

The discussion of the intimate connection between granite and orogeny is illuminating and fascinating. Wegmann's ideas of migmatization and granitization of the lower levels of the crust are engagingly described and contrasted with Suess's *Intrusions tektonik:* "There is no simple fusion at the depths of the geosynclines, the phenomenon is larger and more complex. The depths of the earth are in constant evolution as is the surface, and orogeny simply represents the surface manifestations of this evolution."

Granitic volcanism and granitic plutonism are often, but not necessarily, connected. What does it mean? Is it accidental or caused by laws of fundamental importance?

The granite in the crust of the earth is the subject of another chapter of outstanding interest. Granite forms in the crust by refusion (=migmatization), and by differentiation of gabbroic magma at great depth. "Granitization" is one of the principal functions of the crust because by far the largest parts of the crust are granitic. Granitization is not uniform and continuous; it works in cycles of increasing intensity. Because the nature of the deeper shells of our planet is inadequately known, the limits and the full significance of granitization cannot be stated.

The ore deposits of granitic relation (including uranium and thorium ores) are classified, described, and their mode of origin is interestingly discussed.

The last chapter is concerned with the problems of the formation and the *mise en place* of granite. Although the metamorphic (or metasomatic) mode of origin is given serious consideration on the basis of recent writings by Perrin, Roubault, and Backlund, the author concludes that granite passed from a liquid into a solid state for the following reasons:

1. Temperature of formation is relatively high.
2. Order of crystallization is parallel to Bowen's reaction series.
3. Apophyses and veins of granite occur in adjacent rocks.
4. Granite grades into other massive rocks like syenites, diorites, and others of known magmatic origin.

5. Inclusions of the adjacent rock occur "swimming" in granite.
6. Intrusive massive granites sharply cut sedimentary layers.
7. There is evidence of granite formed by differentiation of a basic magma (Sudbury).

But the author emphasizes that the granitic liquid need not always be an ordinary magma; he believes in an "ichor" whose intimate (mechanical?) mixing with the pre-existing rocks gives birth to migmatites and subsequently to granites. Thus granite is often believed to have formed *in situ* (but the author does not enter into the problem of the volume relations). The granitization of a certain part of the earth's crust is partly an alkalinization, partly a homogenization, the result being the individualization of a granitic magma.

This book is an excellent treatise of the geology of granite; it should be read with equal profit by both the student and the professional geologist. It treats the subject clearly, objectively, and logically and sets up an excellent standard for a textbook. It is stimulating and contains a wealth of information. No doubt the geology of granite is one of the most fundamental subjects in geology; it will expand the domain of the geological sciences which are now too strictly limited to the surface of the earth.

The best praise I can give the book is that the "pontifs" (Bowen, 1947, p. 265) will denounce it as radical and the "soaks" will find it reactionary.

## REFERENCES CITED

BACKLUND, H. G. (1936) Der Magmaaufstieg in Faltengebirgen: Comm. géol. Finlande Bull. 115, pp. 293–347.

——— (1938) The problems of the rapakivi granites: Jour. Geology, vol. 46, pp. 339–396.

——— (1943) Einblicke in das geologische Geschehen des Präkambriums: Geol. Rundschau, vol. 34, pp. 79–148.

——— (1946) The granitization problem: Geol. Mag., vol. 83, pp. 105–117.

BARTH, T. F. W.; CORRENS, C. W.; and ESKOLA, P. (1939) Die Entstehung der Gesteine: Ein Lehrbuch der Petrogenese, Berlin.

BOWEN, N. L. (1947) Magmas: Geol. Soc. America Bull. 58, pp. 263–279.

BRAY, J. M. (1942) Spectroscopic distribution of minor elements in igneous rocks from Jamestown, Colorado: *ibid.*, Bull. 53, pp. 705–814.

GROUT, F. F. (1941) Formation of igneous-looking rocks by metasomatism: a critical review and suggested research: *ibid.*, Bull. 52, pp. 1525–1575.

PERRIN, R., and ROUBAULT, M. (1937) Les Réactions à l'état solide et la géologie. Bull. Serv. Carte Géol. Algérie 5e sér. Petrographie. No. 1.

RASTALL, R. H. (1945) The granite problem: Geol. Mag., vol. 82, pp. 19–30.

READ, H. H. (1943) Meditations on granite I: Geol. Assoc. London Proc., vol. 54, pp. 45–85.

——— (1944) Meditations on granite II: *ibid.*, vol. 55, pp. 45–93.

REINHARD, M. (1934–35). Über Gesteinsmetamorphose in den Alpen: Jaarb. Mijnbouw Vereen, Delft.

REYNOLDS, D. L. (1947) The granite controversy: Geol. Mag., vol. 84, pp. 209–223.

RITTMAN, A. (1942–45) Le Temperature nella crosta terrestre e l'orogenesi: Rend. R. accad. sci. Napoli, vol. 13.

SAHAMA, TH. G. (1945*a*) On the geochemistry of the east Fennoscandian rapakivi granites: Soc. géol. Finlande Compte rendu, vol. 18, pp. 15–67.

——— (1945*b*) Spurenelemente der Gesteine im südlichen Finnisch-Lappland: Comm. géol. Finlande Bull. 135.

SANDELL, E. B. (1943) The rare metallic constituents of some American igneous rocks: Jour. Geology, vol. 51, pp. 99–189.

SHIMER, J. A. (1943) Spectrographic analysis of New England granites and pegmatites: Geol. Soc. America Bull. 54, pp. 1049–1066.

WEGMANN, C. E. (1935) Zur Deutung der Migmatite: Geol. Rundschau, vol. 26.

——— and KRANCK, E. H. (1931) Beiträge zur Kenntnis der Svecofenniden in Finland: Comm. géol. Finlande Bull. 89.

# REVIEWS

*Igneous Minerals and Rocks*. By ERNEST E. WAHLSTROM. New York: John Wiley & Sons, 1947. Pp. vii+367; figs. 135; tables 19. $5.50.

The sequence in which the author presents his material is peculiar. The leading principle, if any, appears to be to make it the opposite of the way nature does it. It is preposterous to discuss the minerals serpentine, prehnite, chlorite-mica and olivine in the order stated and then switch back to zoisite. In like manner the author discusses the rocks in an inverted order. The book is purely descriptive and pays little attention to genetic relations; but that is no excuse for starting with granite, followed by pegmatite, adamellite, tonalite, and quartz gabbro in the order named. Because nature does this, generally speaking, in the opposite direction, it is obsolete to follow the German taxonomers of the last century. The fact that they started it is no reason for our sticking to it.

The book is spotty. While a good description of the optical determination of the plagioclases is offered (although the important high-temperature plagioclase series of Koehler and Tertsch [1942] is neglected), the pyroxenes and the olivines are in this respect treated as stepchildren. The book either should or should not give the necessary data for optical determination of the most important mineral series. I think it should, and I think, furthermore, that it could have been presented without any expansion of the text. The alkali feldspars are poorly treated. Orthoclase, $(K,Na)AlSi_3O_8$, is shown to exhibit variations in the extinction angles but not in the refractive indices. Microcline, with the same formula, exhibits variations in the refractive indices but not in the extinction angles. And sanidine, again of the same formula, differs from both in showing variations of both these properties. Since the alkali feldspars form one of the most important mineral series in nature, it is essential to present them correctly. The student by reading this book (unless he has a teacher who knows better) will become confused and completely fail to grasp the significance of any of these important minerals.

Many minerals are presented by a definite formula, although great variations are claimed in the optical properties, without any reason being given for this peculiar behavior. This is always confusing, but it is particularly so where extended series of solid solutions are treated. For example, data on tremolite and actinolite are presented as follows:

$$Ca_2Mg_5Si_8O_{22}(OH)_2, \qquad \alpha = 1.599\text{–}1.613,$$
$$Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2, \qquad \alpha = 1.614\text{–}1.628.$$

Apparently, the pure magnesium tremolite was able to show just as large optical variations as does the mixed crystal series. The upper limit of index $\beta$ in tremolite is given as 1.671, which certainly must be a misprint. But I cannot even guess which figure the author had in mind, for the pure Mg-tremolite should show no variation at all.

Antigorite, serpophite, and chrysotile are shown to have different optical properties but exactly the same chemical formulas. The student must necessarily conclude that he has a case of polymorphism, like that of andalusite-sillimanite, the formula presentation of which is strictly analogous. Talc is correctly given as $Mg_3Si_4O_{10}(OH)_2$; but antigorite, serpophite, and chrysotile are not allowed to have any (OH)-groups in their formulas. Prehnite has $(SiO_4)$-groups but no (OH)-groups; epidote, however, has both $(SiO_4)$-groups and (OH)-groups. Inconsistencies like these should *never* occur in a textbook.

Modern data are often neglected. Larsen's lectures that Wahlstrom heard before the war are not necessarily modern in 1947: biotite is not categorically monoclinic—hexagonal and trigonal members are known (Hendricks, 1939). On the other hand, the "modern" ideas about the interchange of Al and Si are given a mystical power in, e.g., nepheline, which is presented as $(Na,K)(Al,Si)_2O_4$. I do not see how Al:Si can differ from unity in this formula. $(K,Na,Ca)_{0.8-1}(Al,Si)_2O_4$ would be correct.

Cordierite can no longer be presented as $(Mg,Fe,Mn)_2(Al,Fe)_4Si_5O_{18} \cdot H_2O$. Certainly, this formula cannot account for the great variation claimed for the indices of refraction ($\beta$ = 1.538–1.597) figures, which are also given in the Determinative Tables. The variation in the in-

dices are mainly due to "stuffing" of the lattice with alkali ions. Cordierite, like beryl, has a structure containing $(Si_6O_{18})^{12-}$-rings; no water is present in the formula; but both water and alkalies can be stuffed into the hole in the center of the rings. Halite is said to form through direct normal crystallization of magmas (p. 186). Wahlstrom seems to have been in a great hurry: there are unnecessary repetitions; a poor index that does not give reference to all places where minerals are listed; misspellings like trach*i*te, luxul*l*ianite, etc.

It is depressing to find a detailed description, ten pages long, of the calculation of the norm without any mentioning of Niggli's important introduction of the so-called "molecular norm" (1936). The weight norm of the CIPW-system is obsolete, and in my opinion it should never be used in its original form. Maybe most American petrologists do not know this, but a man writing a textbook has a duty to know.

T. F. W. B.

---

*Rocks and Rock Minerals.* By LOUIS V. PIRSSON and ADOLPH KNOPF. 3d ed. New York: John Wiley & Sons, 1947. Pp. vii+349; figs. 72; tables 50; pls. 36. $4.00.

The original first edition by Pirsson has excellent features from which students and teacher can fully profit. Obsolete statements and those not in harmony with modern concepts can be pointed out for the students and thus become of pedagogical value. The 1947 edition, however, must be measured by modern standards. It falls far short.

Poor judgment is used in the sequence in which the minerals are presented: mica, pyroxene, hornblende(!) . . . quartz . . . ! The sequence should be either genetic or chemical. Here no principle is followed. The whole is a confused mixture. Pyrite and apatite are placed under "Oxides." Chlorite and zeolite are placed under "Hydrous Silicates," but mica is placed with feldspar, etc., under "Silicates."

Under feldspar, perthite is *not* mentioned, although it is probably the most common feldspar mixture in rocks.

The presentation of the chemical composition of minerals is a hopeless mixture of right and wrong, old and new. The discussion of polysilicic acid, etc., is obsolete.

Amphiboles should not, "like pyroxenes, be considered to be metasilicate," and their compositions are *not* "too complex to be represented by simple formulas." Why does epidote contain a (OH)-group and a $(SiO_4)$-group, while mica has no (OH)-group, and idocrase has a (OH)-group but no $(SiO_4)$-group?

The formula for staurolite is radically wrong. To the Freshman it might be presented as a sillimanite structure interpenetrated by sheets of iron hydroxides:

$$\left\{\begin{array}{l} Al_2SiO_5 \\ Fe(OH)_2 \\ Al_2SiO_5 \end{array}\right.$$

In the discussion of the chemistry of tourmaline, the terms $W$ and $X$ are suddenly used. If all the incoherent discussions of the mineral chemistry had been based on such conceptions from the very beginning, the book would have been much easier for the student to digest. Now great confusion prevails: there is no correspondence between related minerals; the formula for serpentine is presented differently from that of clinochlore, while the general formula of chlorite is altogether neglected; there is no formal connection between the various zeolites; and such monsters as $Na_4(AlCl)Al_2(SiO_4)_3$ for sodalite make it impossible for the student to get any understanding of mineral chemistry.

In the petrographic section, grains of magnetite enclosed in biotite are discussed: "They are evidently older than the biotite, because they are enclosed by it." This is far from true. By implication, albite lamellae should be older than microcline because they are enclosed by it! A flat statement like: "Igneous rocks show in general this order of crystallization: first, the oxides of iron, then ferromagnesian minerals, then plagioclase feldspars, then alkali feldspars (and feldspathoids), and lastly quartz" is both untrue and harmful because it violates the fundamental laws of the physicochemistry of silicate melts. Both through the text and through graphical illustrations it is implied that the ferromagnesian minerals all cease to crystallize before the final stages of the crystallization history is reached. Yet no phase can cease to crystallize in a cooling system unless it is replaced by its reaction product. But plagioclase and quartz are clearly not reaction products of ferromagnesian minerals.

Knopf uses the old illustrations which, by now, are worn and poor; there is an almost complete lack of new references; coal has been forgotten in the classification of sedimentary rock. But these are trifles; one might call them misprints, and no book is free from misprints.

T. F. W. B.

*An Introduction to Paleobotany*. By CHESTER A. ARNOLD. New York: McGraw-Hill Book Co., 1947. Pp. ix+415; figs. 187. $5.50.

This new paleobotany text presents a systematic study of fossil plants at a level suitable to elementary and intermediate college classes in botany, paleobotany, and general paleontology. It fills a real need, inasmuch as it is the only text to be published in a number of years covering both the nonvascular and the vascular plants. The organization is excellent, and the space devoted to the various groups of plants is well distributed. The principal criticism of the book concerns the style, which is labored, in places trite, and burdened with cumbersome phraseology. This must reduce the effectiveness of the book as a text.

The approach to the subject is basically botanical, but part of the Introduction and one of the final chapters consider the geological aspects of paleobotany. In keeping with the purpose of the book, which is to present an up-to-date text on fossil plants, the classification is taken, with slight modifications, from J. A. Eames. Ten of the sixteen chapters comprise a systematic treatment of fossil plants within the framework of this classification. The second chapter is devoted to an extended, if somewhat elementary, account of the fossilization of plant materials. The third chapter deals with the nonvascular plants in summary style. The last two chapters give brief accounts of plants and environments and the systematics of paleobotany. Each chapter includes a well-selected list of references. Scientific terminology is introduced and freely used throughout the text. Although explanations of terms are included in the text, a glossary would prove of great use to the nonbotanical student.

Each chapter which deals with a major group of plants includes one or more sections which analyze the evolution of the group and its relationships to other related stocks. These analyses may be subject to criticism by botanists and paleobotanists, since they are necessarily somewhat subjective and since it is impossible in a text of this length to consider all evidence and all hypotheses. Nevertheless, the treatment of phylogeny is good and, for the most part, adequate for a text at this level. Only occasionally does dogma, so apparent in many texts, mask the true scope of thought on phylogeny.

Many of the illustrations are new, and few have appeared in other texts. Photographs are well selected and well described, but their clarity is reduced by the use of a rather coarse screen. Drawings are largely diagrammatic and executed with a simple line-and-stipple technique. Some are excellent, and others are rather crude, but all are easily understood.

This text is certain to be of great use to teachers in paleobotany and related fields. It has much to recommend it and relatively few serious defects. It may aid materially in reviving interest in the field of paleobotany, which in recent years has tended to decline in all but a few centers of active research.

E. C. O.

---

*Outlines of Paleontology*. By H. H. SWINNERTON. 3d ed. London: Edward Arnold & Co., 1947. Pp. x+393; figs. 367. $8.00.

In the Preface to the third edition of *Outlines of Paleontology* it is stated that the book has been thoroughly revised. This is something of an overstatement, since much of the text has been changed little or not at all and the illustrations with very few exceptions are the same. It is evident, however, that all sections have been carefully studied, for minor changes in phrasing and terminology occur throughout; and it may be supposed that the volume includes the full scope of changes which the author felt necessary.

Principal modifications have been designed to modernize the book by changing terminology, including new facts and trends of thought, and by changing or eliminating concepts which are no longer considered valid. The greatest changes are found in the sections concerning the Protozoa, Cephalopoda, Trilobita, and fishes. In each instance there has been considerable improvement. A tendency to place greater emphasis on ontogenetic and evolutionary development is evident in each of the revised sections. There has been less recasting of other sections, but those dealing with sponges, alcyonarian corals, and amphibians have been modified by additions or changes.

The changes have been distinctly conservative, which, in general, seems advisable in a standard textbook. The revision does not, however, fully reflect the amount of work and the changes in viewpoint which have taken place since 1930. Some sections, such as that on the Vermes, which were inadequate in the first and second editions, have not been modified materially. The third edition is distinctly superior to the second; but it would have been possible to improve it even more, had the author seen fit

to make more drastic changes in a number of sections.

E. C. O.

---

*Principles of Micropaleontology.* By MARTIN F. GLAESSNER. U.S. ed. New York: John Wiley & Sons, 1947. Pp. xvi+296, figs. 64, tables 7, pls. 14. $6.00.

The American edition of this work which was originally published in Australia in 1945, has now appeared and should prove extremely useful to micropaleontologists and particularly to students. As the name suggests, it reviews the entire specialized field of micropaleontology and includes much practical information not readily available elsewhere. The subject matter is presented in three parts and an Appendix.

Part I begins with a brief introductory chapter on the origin and present status of micropaleontology and continues with a classification and brief review of microfossils other than Foraminifera. This is the only special consideration given to the other fossils, and some parts are much too brief. For example, less than five pages are devoted to ostracodes and less than three pages (exclusive of text figures) to conodonts and scolecodonts combined. This part is concluded by a chapter on collecting, preparing, and studying microfossils, which is excellent.

Part II, devoted to the Foraminifera, constitutes about half the book. It begins with accounts of the life-history and general morphology of these organisms which are well organized, up to date, and, at the same time, brief but complete. The next chapter (over 100 pp.) is concerned with classification. The principles of classification are clearly explained, and previous classifications are compared. Then follows a systematic account of nearly 300 genera and subgenera arranged in 37 families. This section is very similar to the presentation of like material in Cushman's and in Galloway's well-known works but is less complete. Glaessner's classification in general follows Cushman's but is somewhat simplified and considerably altered in parts. A noteworthy feature is the grouping of families in 7 superfamilies, whose presumed phylogenies are discussed. Generic descriptions are brief but clear and adequate. The type species and stratigraphic range of each is given. Of particular importance are the excellent illustrated explanations of structural complexities which characterize certain groups. Unfortunately, this section as a whole is inadequately illustrated, and there are no cross-references between eight plates of Foraminifera and the text. A compensating feature, however, is a series of extremely useful tables which contrast the main morphological characters of more than 100 of the more important genera arranged by families. This part is concluded by a chapter on the paleoecology of both benthonic and pelagic Foraminifera.

The third part begins with a summary of the known stratigraphic occurrences of all types of microfossils and is continued by a chapter on the principles and application of stratigraphic correlation with special reference to microfossils and by another on the application of micropaleontology to petroleum exploration in various parts of the Eastern Hemisphere. The latter appears to be out of place in this book and is so general that it has little importance except for the presentation of some little-known historical facts and a very broad and incomplete tracing of certain contemporaneous ecological zones. Finally, a short chapter is devoted to a list of equipment needed by micropaleontologists and the suggested layout of a laboratory. This is much more closely related to the last chapter of Part I than to the foregoing portions of Part III.

The Appendix consists of a tabular synopsis of Glaessner's classification of the Foraminifera, with the stratigraphic range of each genus and the so-called "larger" Foraminifera distinguished; a table showing the equivalency of his families to Cushman's; and a bibliography of 30 pages arranged by subjects according to chapters in the text. Separate indexes are provided for subjects discussed, fossils mentioned or described, and authors to whom reference is made. The volume is concluded with a table showing the standard European stratigraphic scale.

This book is obviously the work of a practical micropaleontologist of wide experience. Some of the more general sections, like those on ecology and the principles of classification and stratigraphic correlation, reflect a sensible modern viewpoint and an understanding of the essential biological background of paleontology. Their application is not bounded by the limits of micropaleontology.

This book is less a handbook or reference work than is either Cushman's or Galloway's volume. With some supplementing of the nonforaminiferal microfossils, it will serve admirably as a very practical textbook of micropaleontology.

J. MARVIN WELLER

*When the Earth Quakes*. By James B. Macelwane, S.J. Milwaukee: Bruce Pub. Co., 1947. Pp. 288; figs. 251. $5.00.

There are various ways of appraising a book, one of which is for the reviewer to place himself, figuratively, in the position of the reader to whom the book is addressed and then determine to what extent the writer has succeeded in telling him what he wishes to know. The present book is one volume of a "Science and Culture Series" edited by Joseph Husslein, S.J., Ph.D., and is intended principally for the instruction of the lay reader on the subject of earthquakes and related phenomena. Putting himself in the position of the lay reader, the reviewer is obliged to disregard any previous knowledge he may possess about earthquakes and earth science in general, other than newspaper, magazine, and newsreel accounts and other such general types of information. Thus divested, he wishes to know what the directly observable facts concerning earthquakes are and, finally, what indirect conclusions may be deduced from these facts and *how the deductions are made*. If he is critical, he will refuse to accept assertions in the form of unsupported inferences.

Reading the book from this point of view, one is pleased to find the first three chapters, including eighty-five figures of photographs and maps, devoted to the description of what actually has been observed in earthquakes, with a brief outline of the principal facts pertaining to some two dozen separate earthquakes. In chapter iii, on "Kinds of Earthquakes," however, the argument becomes a little more difficult to follow when earthquakes are divided by causes into the classes of "tectonic," "volcanic," and "plutonic" earthquakes. The first two classes are substantiated fairly conclusively by the evidence presented, but the last, the "plutonic" type, is supported only by the authority of the author's assertion.

The next four chapters (iv, v, vi, and vii), entitled "Why Earthquakes?", "When and Where?", "Field Study of Earthquakes," and "Sea Quakes and Seismic Sea Waves," are moderately straightforward. The first two deal mostly with direct geological evidence, and the third discusses principally the Abridged Wood-Neumann Scale of seismic intensities and its use in seismic mapping by means of isoseismal lines.

There is, however, in the third of these chapters a logical inversion in a section devoted to the nature of earthquake motion, wherein *P*-waves and *S*-waves are discussed and are asserted to be, respectively, dilatational and transverse elastic waves, notwithstanding the fact that *P*- and *S*-waves are unknown to the reader until he is introduced to seismograms in a later chapter. Also no evidence is presented in support of the assertions as to the nature of the waves.

The remainder of the book, chapters ix–xvii inclusive, is devoted to various aspects of instrumental seismology, including earthquake, engineering, and prospective seismology, with a final chapter on microseisms and their relation to storms at sea. The treatise on earthquake seismographs is very good; but, when the reader expects to learn how seismograms are employed to decipher the nature of the interior of the earth or to determine the focal depth of plutonic earthquakes, he is disappointed. Seismograms showing *P*- and *S*-waves are reproduced, and it is asserted, but with no supporting evidence, that these travel *through* the earth (whereas another type of wave is propagated along the surface).

The chapters on engineering applications of seismology and a brief account of prospecting methods are appropriately done for present purposes; also that on microseisms. An excellent sheet of ten seismic prospecting records is presented in Figure 220, showing the correlation of about five different reflection horizons.

Appendixes containing a brief outline of geology, an excellent seismic bibliography of 105 titles, and a glossary of terms are added, which greatly increase the usefulness of the book.

From the foregoing account it will be inferred that, on the whole, this book achieves its objective rather well. There are, however, as indicated, several serious lapses of logical consistency, and it is regretted that a better job was not done in presenting at least the simplest deductions from seismic records.

It is also regrettable that a shadow should have been cast upon the scientific integrity of the book by its editor, who in his Preface, after reciting a case in which an ignorant Italian workman attributes earthquakes to a "scourge of God," proceeds to assure the reader: ". . . he [the workman] was correct in supposing that these natural phenomena occur by the disposition of God's Providence."

M. King Hubbert

---

*Some Structural Features of the Intrusions in the Iron Springs District*. By J. H. Mackin. ("Guidebooks of the Geology of Utah," No.

2.) Salt Lake City: Utah Geological Society, 1947. Pp. 62; figs. 12.

During the war years the steel industry of Utah, Colorado, and California depended largely on the iron-ore deposits of Iron Springs, Utah, located about 220 miles south-southwest of Salt Lake City. This paper presents a stimulating preliminary interpretation of the origin of the deposits, which were investigated by Mackin and his assistants under the auspices of the United States Geological Survey, aided by drilling and trenching operations by the United States Bureau of Mines.

Three small intrusions of quartz-monzonite porphyry are lined up in a northeast-southwesterly direction and have domed and faulted contacts with a series of Jurassic (Homestake, "Entrada") and Upper Cretaceous to Eocene (Iron Springs, Claron formations) sediments. At the base of this series is found, in near-by areas of the Colorado Plateau, a gypsum horizon with interbedded limestone and shale, which, in turn, rests on the thick Navajo sandstone. Following a local pre–Iron Springs disturbance that brecciated the Homestake limestone in places and brought about local erosion of the Entrada series, the monzonite magma is believed to have broken through the resistant Navajo sandstone along a zone some distance east of the present exposures, to have spread along the easily yielding gypsum horizon, and to have lifted the overlying strata to form laccolith- or bysmalith-like bodies of monzonite porphyry.

According to Mackin, the origin of the iron ores in the contact zones of the intrusions is closely related to fracturing of the newly consolidated igneous rocks. Thus the western border of the Three Peaks intrusion is penetrated by subvertical joints that strike approximately at right angles to the nearest contact. The fractures are believed to have been caused by distension of the monzonite porphyry at a stage when the magma had already formed a chilled border and cooling was advancing inward. Other fractures dip into the intrusive, striking approximately parallel to the contact. The first fracture system indicates elongation of the consolidating mass in subhorizontal directions, more or less parallel to the strike of the border; the second system records a marginal lengthening of the cooling mass in directions down-dip along the domed roof. Other fractures seem to occupy intermediate positions between these two systems and may curve from the one into the other principal direction.

These conspicuous fractures are commonly coated with magnetite selvages, a fraction of an inch to several feet thick. The author assumes that the expulsion of iron oxide was essentially a deuteric process, comparable to the release of volatiles from cooling basalt along contraction joints, drawing its material from the cooling rock near the joints. Numerous analyses showing deficiency in iron in monzonite porphyry directly adjacent to magnetite selvages are cited in partial support of this theory. Particularly strong doming or expansion in local areas is thought to have facilitated the opening and wide gaping of suitable fractures, whereas shearing between the chilled border and the fractured zone of monzonite next below has inhibited the outward expulsion of emanations. Marginal fault zones, cutting across the intrusion and the wall rocks in places, are believed to have permitted the iron-precipitating emanations to reach the Homestake limestone, in which case replacement ore bodies of prodigious size were formed.

As the author emphasizes, several points in this theory of the origin of the ore deposits have not yet been determined quantitatively. Even so, the structural study of the district illuminates significantly the mechanism that permits transfer of iron-precipitating emanations from the interior of an intrusion through an intensely fractured outer zone, leading under favorable conditions to large-scale replacement ore bodies in carbonate rocks. Structural factors that facilitate such transfer and others that impede or prevent it are clearly identified; the usefulness of detailed areal structure surveys for better understanding of problems of ore concentration is convincingly shown; and guiding principles for locating additional favorable areas of ore concentration are pointed out. The forthcoming full report on the area, now being prepared by the Geological Survey, should prove interesting reading.

R. Balk

---

*The Pegmatites of Central Nigeria.* By R. Jacobson and J. S. Webb. (*Geol. Survey of Nigeria Bull. 17* [1946].) Pp. 61 (small 4to); pls. 6; figs. 10; folded maps 4. 10/6 from The Technical Bookshop, 724 Salisbury House, London Wall, London.

Although this report is a product of "war research," some two years were devoted to its

preparation. Several hundred pegmatites (a few of which are briefly described in an appendix), and some four hundred thin sections were examined. Sixteen photomicrographs, showing some interesting details, and a generalized paragenetic diagram are included in the report. As in most studies of pegmatites, many of the conclusions reached are not sufficiently buttressed by a logical chain of evidence; nevertheless, this report is a publication of much more than average value.

The Nigerian granites are divided into "Older" and "Younger." The former embraces a variety of pre-Cambrian granites which probably belong to several different periods of plutonic activity; these are genetically related to the pegmatites. The younger granite occurs in north-central Nigeria; it contains biotite and riebeckite and has given rise to the major (alluvial) tin deposits.

The pegmatites occur as fairly tabular dikes up to a mile or more in length and 30 feet or more in width; the more highly albitized ones have gentle dips and show pinch and swell structure. Simple pegmatites occur in the granites; albitized pegmatites carrying cassiterite, columbite, etc., are found in the surrounding metamorphic rocks and are not conformable with the foliation of the host rock; quartz-tourmaline veins, locally (fig. 4) closely related to the pegmatites, may be conformable with the foliation of the country rock.

The first stage in the development of the pegmatites is generally characterized by perthitic microcline and quartz, and a slightly earlier quartz-muscovite ($K_2O:Na_2O = 8:1$) border zone. Minerals of this stage were followed by the deposition of albite, some of which was injected among the earlier pegmatite minerals to form a fine-grained, sugary, pseudo-aplitic rock. Part of the albite is coarse cleavelandite, with which is associated columbite, cassiterite, muscovite ($K_2O:Na_2O = 4:1$), phosphates, etc. The deposition of the albite and associated minerals was succeeded by formation of the quartz-tourmaline veins; minor kaolinization and chloritization were the last processes operative.

The early formation of small poikilitic crystals of cleavelandite, as advocated by Andersen and the reviewer, is not recognized by Jacobson and Webb. Such early-formed cleavelandite, however, seems to be shown in plate II, figures 3 and 4. The authors have presented convincing evidence that the albitization process is characterized in part by replacement of the microcline. The nature of the wall rock has little or no effect on the mineralogy of the pegmatites.

D. J. F.

---

*Le Congo physique*. By MAURICE ROBERT. 3d ed. Liége: H. Vaillant-Carmanne, S.A., 1946. Pp. 449; figs. 70; pls. 29.

The first edition of this authoritative work was prepared under the auspices of the Comité Spécial du Katanga in 1919 and the second in 1941. While no great time has elapsed since the appearance of the second edition, there have been some notable new contributions to our knowledge of central Africa, which have been utilized by the author to bring the present volume up to date.

In this treatment the Congo geology has been, to a helpful extent, tied in with that of South Africa, enabling the reader to get a broad general picture, as well as the regional details. The mineral deposits are of great importance, and some of them of world-wide interest. Full treatment is also given to the evolution of the surface relief, together with the hydrographic system, and to the climate. A long chapter is devoted to biogeography.

R. T. C.

---

*A Yanqui in Patagonia*. By BAILEY WILLIS. Stanford University: Stanford University Press, 1947. Pp. 152; figs. 32. $3.00.

The Yanqui is Bailey Willis. The ever youthful nonagenarian has been persuaded by appreciative friends to write a book of reminiscences of his unusually eventful life. Perhaps these friends urged an autobiography, but the resulting literary production is that only in part; the narrator is too much of an artist to act like an ordinary biographer. Instead, he has painted only certain parts of the whole picture, giving them a vividness and charm peculiarly his own.

Part I, "Before Patagonia," introduces the Yanqui. His graduation from the Columbia School of Mines coincided with the establishment of the United States Geological Survey, and his professor of geology recommended him to Clarence King. King turned him over to Pumpelly, who sent him on a solitary ride of 600 miles through the valleys and mountains of the south to find the "ore banks" from which the Confederates made iron. Next he was directed to run down reports of iron ores north of Lake Superior, which he did in a long trip with Ojibway Indians that began in birchbark

canoes and ended on snowshoes. Two years later, Pumpelly assigned him the task of surveying coal lands, believed to occur widely in the dense forests between Puget Sound and the Cascades, for the Northern Pacific Railway, then being built. Some of the most interesting pages of the book tell the story of this formidable undertaking.

Nearly thirty years ago, the reviewer was impressed by the following in Raphael Pumpelly's *Reminiscences* (vol. II, p. 627) (the scene was Tacoma):

> There was a trial of a ship's captain for extreme brutality. The captain swore it was the only way to manage his crew. The prosecuting attorney hurled at him: "Do you mean to say that you and your mate can't manage a half-dozen sailors when the boy Willis up in the woods, easily and alone, controls ten times as many of the toughest scoundrels in the country?"

Pumpelly's reference is readily understood, on reading the present book, as just a typical reflection of Willis' surpassing skill in dealing with men, whether though miners, lumberjacks, hostile Indians, or smooth Argentine bureaucrats. The stories of these dealings make delightful reading.

A single short chapter, "The United States Geological Survey," is all there is to represent a major hiatus between the author's early exploits and the Argentine project of 1910–1914. But it does give some revealing sidelights, particularly on Powell and Gilbert.

Part II, "Patagonia," is the greater part of the book. Don Ezequiel Ramos-Mexia, minister of public works of Argentina, was facing grave difficulties in prosecuting his plans for the development of lands still belonging to the nation. Water was the great need and, since semiarid areas in the United States had been made productive by drilling artesian wells and building railroads, Ramos-Mexia had built many miles of railroads and caused many wells to be sunk, but he had not found water. An American geologist to head a "Comision de Estudios Hidrologicos" was his remaining hope, and he contracted Bailey Willis.

Willis went at the job with an enthusiastic party of young Americans, but their geological survey proved artesian water to be a vain hope; a dam-and-reservoir project was the best they could offer. The Minister then put them to surveying the best route for a southern transcontinental railway across the Andes. Their extensive explorations Willis enlivens with tales of the Conquistadores, the city of Los Cesares, the cave man of Lost Hope Fiord, and other delightful bits of local color. Beautiful Lago Nahuel Huapi was the focus of many of the Yanqui's activities, and in planning a future development of this area with railroads, an industrial city at the east end of the lake, and a national park beyond, he made some notable contributions, indifferently appreciated by Argentine officialdom.

R. T. C.

---

*On Seismogram Types and Focal Depth of Earthquakes in the North Japan and Manchuria Region.* By EIJO VESANEN. ("Isostatic Institute of the International Association of Geodesy Publications," No. 15.) Helsinki, 1946. Pp. 25; figs. 14.

*On the Gravity Field and the Isostatic Structure of the Earth's Crust in the East Alps.* By PAAVO E. HOLOPAINEN. ("Isostatic Institute of the International Association of Geodesy Publications," No. 16.) Helsinki, 1947. Pp. 94; diagrams 5; maps 2. $1.50.

PUBLICATION NO. 15

The first is a sequel to an earlier paper of 244 pages published by Vesanen in 1942, on the type-evaluation of seismograms. The reviewer has not seen the earlier paper, but it evidently presented the thesis that there was a recognizable character difference in seismograms from shallow and deep foci and, furthermore, that the depth of transition was not constant from area to area.

The present paper is an application of that thesis to the earthquakes of northern Japan and Manchuria, recorded at Helsinki between January 1, 1925, and June 30, 1940. A complete table is given for these quakes, showing, besides their computed focal depths, their seismogram types, NJa 1, NJa 2, and NJa 3, based upon the character of the *P*- and *S*-waves. These three character types correspond, in the author's judgment, to three different physical layers in the order of increasing depth, in which the different types of earthquakes originate.

On a map showing the location of the epicenters, the shallow-focus quakes (NJa 1 type) are clustered mostly in a linear belt along the continental slope between Honsyu and the foredeep to the east. The shocks of intermediate depths occur somewhat westward of this zone, and the deep shocks still farther to the west.

Although photographs of the $P$- and $S$-records of a number of earthquakes are shown, it is not clear to the reviewer from the present paper what the diagnostic features of the different types are. It also appears that a strong element of personal judgment is involved in the classification. This may account for one of the principal conclusions of the paper. In certain quakes the author has found that $P$ has a shallow character, whereas the character of $S$ is intermediate. From this he concludes that the $P$- and $S$-waves have originated *in different layers.*

To the several seismologists and physicists with whom this hypothesis has been discussed it appears absurd and physically impossible. One questions, therefore, whether a better explanation may not be found in the unreliability of the criteria and the fallibility of subjective judgment.

PUBLICATION NO. 16

The second of these papers is an analysis of the gravity data of the region of the eastern Alps. This area comprises roughly the quadrangle bounded by the 45th and 49th parallels of latitude and the 9th and 17th east meridians.

The gravity data for the area consist of 301 stations, all but 91 of which were measured prior to 1894. The most recent stations were the 91 measured between 1910 and 1932. Intercomparison between different surveys permitted the elimination of all obviously "wild" values, some of which were apparently in error by as much as 50–60 milligals. For the remaining stations, free-air, Bouguer, and local Airy-Heiskanen anomalies were computed, the latter for each of the seacoast crustal thicknesses of $T = 20$, 30, 40, and 60 km. In certain selected areas, tests of regional versus local compensation were also made by computing regional anomalies for various radii of assumed regional compensation, using the method of Vening Meinesz.

Maps of iso-anomaly lines are shown for the Bouguer anomalies and also for the Airy-Heiskanen anomalies for $T = 20$ km. These show definite axial trends parallel to the tectonic axis of the mountain range. Across this axis several profiles are plotted, showing, in addition to the topography, the profiles of free-air, Bouguer, and the two or three Airy-Heiskanen anomalies of least amplitude.

The Bouguer anomalies, as usual, are the largest, showing a broad negative trough of a depth of $-175$ milligals along the axis of the range. Next comes the free-air anomaly, also dominantly a negative trough, and, finally, the various isostatic anomalies, the smaller of which show an amplitude of variation of from about $-25$ to $+100$ milligals, with alternate ridges and troughs parallel to the axis of the range.

In addition to the graphical representation of the data, they also were treated statistically. The customary sums of the squares of the various types of anomalies were computed and tabulated. Also a new test was introduced, which the author thinks—and the reviewer agrees—is more diagnostic. This was a statistical correlation of each class of anomaly with the difference between the elevation of the station and that of the surrounding region. The best anomaly by this criterion would be the one having the smallest elevation correlation.

For the region as a whole the best results were given by the Airy-Heiskanen anomaly for $T = 20$ km. and local compensation. In certain parts of the region better results were given by $T = 30$ km. and in others by regional compensation to a few tens of kilometers. Even so, there remained belts of positive and negative anomalies of 40–50 milligals. These the author explains by the buckling hypothesis of Vening Meinesz. One negative anomaly of 40–50 milligals in the northern part of the area was thought to be produced by light-sediments some 3–10 km. thick.

While this is a very able study, it seems to the reviewer that there are two conflicting interests which might be better served by separate methods of analysis. The first of these is a test of the isostatic hypothesis; the second is an inquiry into the actual distribution of anomalous masses. The methods used here and in other studies of this kind are explicitly designed to test the hypothesis of isostasy; what the reviewer would like to see would be comparable methods uncontaminated by any kind of the apriori hypothesis and designed expressly for the purpose of yielding the maximum information as to the magnitude and disposition of anomalous masses.

He also would like to see pendulum data supplemented and supplanted by gravity-meter data accurate to 0.1 milligal.

M. KING HUBBERT

---

*Two Problems of Marine Geology: Atolls and Canyons.* By PH. H. KUENEN. (*K. Akad. Wetensch. Amsterdam Verh., Afd. Naturk.*, Tweede Sectie, Vol. 43, No. 3.) Amsterdam, 1947. Pp. 67.

This publication, based on two lectures delivered at London University in November, 1946, is a well-considered appraisal of two outstanding geological problems.

In the section on atolls, the results of recent borings on the island of Maratoca, northeast of Borneo, are described for the first time. The deepest boring penetrated coral limestone, coral detritus, and marly limestone to a depth of 429 meters and is believed to be evidence of subsidence of at least 500 meters during coral upgrowth. Following a review of the principal theories of origin, a theory of *glacially controlled subsidence* is presented. According to this theory, preglacial atolls and barrier reefs were formed by subsidence; during glacial stages platforms were eroded at glacial sea level; and, with postglacial rise in sea level, the corals grew upward along the margins of the platforms. It is emphasized that important erosion was accompanied by chemical attack, which may have permitted continuous coral growth during the glacial stages.

The section on submarine canyons presents a review and analysis of critical data and principal hypotheses of origin. Turbidity currents, due mainly to wave turbulence and partly to slumping, are favored as the most reasonable explanation offered so far. Tank experiments by the author and certain features of the canyons themselves are presented in support of this view. It is recognized that the major objection to the hypothesis is the hardness of certain rocks which had to be eroded, notably in the California canyons.

L. H.

---

*Field Conference in the Bighorn Basin: Guidebook*. Arranged by the UNIVERSITY OF WYOMING, WYOMING GEOLOGICAL ASSOCIATION, and the YELLOWSTONE-BIGHORN RESEARCH ASSOCIATION. 1947. Pp. 277; figs. 51; pls. 18; maps and sections in pocket. $4.00. Obtainable from the Wyoming Geol. Assoc., P.O. Box 545, Casper, Wyoming.

In August, 1946, the above-named institutions conducted a field conference in the Bighorn Basin of Wyoming to see and discuss the salient features of that exceptionally interesting region, where geologists have been very active in recent years. At this conference, Fellows of the Geological Society of America, participating as special guests, inaugurated a Rocky Mountain Section of the Society. The large guidebook, carefully prepared for the occasion, opens with generalized stratigraphic sections and then proceeds with the detailed road logs for the four days of the conference. For the last two days, seven optional trips were offered; and, finally, since the participants could not get out of the Bighorn Basin without crossing its very significant border structures, they were given "exit road logs" for five alternative routes, together with two short side trips. Accompanying the road logs are concise general statements, local columnar sections, cross sections, sketch maps, and fourteen plates of photographs grouped together for convenience.

Twenty special papers make up the last two-thirds of the volume, devoted primarily to stratigraphy, structural relations, and typical oil fields. It is really invidious to call attention to but two of these individual contributions; but readers who relish radical departures from current interpretations will find such in J. Hoover Mackin's strong contention that "the Middle Rockies are essentially in a first, not a second or third cycle of erosion" and in Walter H. Bucher's conviction that there never was a great "Heart Mountain overthrust" sheet. Mackin concludes that the middle and late Tertiary relief of the Beartooth and Bighorn ranges was essentially a continuation of the relief developed by Laramide movements, decreased, on the one hand, by erosion and, on the other, by aggradation of lowlands, and that the subsummit surface is not an early Tertiary peneplain but is instead a pediment, probably formed much later at not more than 2,000 ± 1,000 feet below the present altitudes of its remnants. Bucher regards the widely scattered, but relatively small, allochthonous limestone masses of the "Heart Mountain overthrust" not as remnants of a continuous orogenic thrust sheet but as individual blocks that have slid down slopes basinward under the action of gravity. Both men have presented much evidence supporting their interpretations.

Conferences of this sort are becoming popular for their effective dissemination of knowledge and their stimulus to further research. Participants gain the most; but the guidebooks, filled with judicously assembled information, can spread the benefits much more widely. They are very handy as books of reference.

R. T. C.

# COMMUNICATIONS AND ANNOUNCEMENTS

## THE NATIONAL RESEARCH COUNCIL HAS APPOINTED A COMMITTEE ON GEOPHYSICS, ADVISORY TO THE OFFICE OF NAVAL RESEARCH

Acting upon a request from Admiral Paul F. Lee, chief of Naval Research, the National Research Council has appointed a committee of active scientific investigators to advise the Geophysics Branch of the Office of Naval Research regarding scientific and related aspects of their research programs. The committee is as follows: Walter H. Newhouse, *chairman;* Harry H. Hess, *vice-chairman;* and Roland F. Beers, Maurice Ewing, Ellis A. Johnson, Lester E. Klimm, William C. Krumbein, William W. Rubey, and J. Frank Schairer.

Dr. R. C. Gibbs, chairman of the Division of Mathematical and Physical Sciences of the National Research Council, acts as administrative adviser to the committee, in collaboration with Dr. Arthur Bevan, chairman of the Division of Geology and Geography.

The committee held its first meeting on January 7 and 8, 1948, in Washington, D.C. As stated by Dr. Roger Revelle, head of the Geophysics Branch of the Office of Naval Research, that branch is charged with the responsibility within the Navy Department of sponsoring basic research in appropriate fields of earth sciences through financial and other support of worthy projects. Functioning within this general framework of responsibility, the committee will, for the present, restrict its consideration to research problems dealing with the crust of the earth and the properties of the earth as a whole.

Since it is part of the policy of the Office of Naval Research to sponsor research in fields not adequately covered by other agencies, the Geophysics Branch, with the committee's concurrence, has established the following objectives for research within the fields covered by the committee:

1. To foster, in co-operation with other agencies, geological, geographical, and geophysical explorations of little-known areas of the earth, such as the islands of the western Pacific and the Arctic and Antarctic. Such exploration may include all aspects of the natural environment and problems of human and economic geography and ethnography, as well as the more limited objectives implied in the terms "geology" and "geophysics."

2. To conduct laboratory and field studies leading to a greater understanding of the properties and processes existing in the outer hundred kilometers of the earth's crust.

3. To develop instruments and techniques for determination of the earth's properties, for example, universal airborne magnetometer equipment.

The Office of Naval Research has adopted the policy of avoiding formulation, detailed direction, and the "farming-out" of projects for basic research, in the belief that maximum progress and results will be realized if projects for investigation grow out of the ideas and interests of the investigators themselves. The committee warmly indorses this policy and will seek at all times, through advice and recommendations, to further its operation to the maximum degree possible. The submission of significant and well-organized research proposals will be helpful in this connection.

For further details consult the Office of Naval Research, Navy Department, Washington 25, D.C., or the National Research Council, 2101 Constitution Avenue, Washington 25, D.C.

## *UNIVERSITY OF KANSAS PALEONTOLOGICAL CONTRIBUTIONS*

The first numbers of a new scientific publication, entitled the *University of Kansas Paleontological Contributions*, have been issued (page size, $8 \times 10\frac{3}{4}$ inches; full-tone plates). Separate series bearing the names of phyla are established for articles dealing with representatives of these phyla; and ultimately title-pages covering a group of related articles will be issued, for use in binding.

Copies of the *Contributions* will be distributed to individual paleontologists on request without charge other than a fee of 25 cents for mailing.

## DRAINAGE ALIGNMENT ON THE SAGANAGA GRANITE

The drainage pattern on igneous rock masses usually shows no evidence of structural control. In a recent paper (1947, p. 356), the writer states that the numerous lakes on the homogeneous Saganaga granite in northeastern Minnesota have no definite alignment. Dr. F. F.

Grout, who has made a careful study of the Saganaga batholith (1936), states that there is a strong relationship of the drainage pattern to a systematic internal structure curved around a dome.[1] There is an "elliptical" arrangement of the lakes which is not closed on the east. The main control, according to Grout, must be the cross-joints and dikes (normal to the flow-lines) and inclusions and schlieren streaks having easily weathered minerals.

REFERENCES CITED

GROUT, F. F. (1936) Structural features of the Saganaga granite of Minnesota-Ontario: Internat. Geol. Cong., Washington, 1933, Rep. 16.

VER STEEG, KARL (1947) The influence of geologic structure on the drainage pattern in northeastern Minnesota: Jour. Geology, vol. 55, pp. 353–361.

KARL VER STEEG

## INTERNATIONAL GEOLOGICAL CONGRESS, XVIII SESSION GREAT BRITAIN, 1948

*General Secretary:* GEOLOGICAL SURVEY AND MUSEUM
EXHIBITION ROAD, LONDON, S.W. 7

29th January, 1948

*Dr. H. R. Aldrich,*
*The Geological Society of America,*
*419 West 117th Street,*
*New York 27, New York*

DEAR DR. ALDRICH,

Thank you very much for your letter of January 7th. Professor Read certainly found his visit and the welcome you gave him most enjoyable and stimulating.

He tells me that he found that certain difficulties in obtaining shipping passages are being experienced by American geologists who wish to join the Congress; and I have received directly one or two complaints and enquiries about the same point. I have taken the matter up with Cooks and with Cunard White Star (who gave us assurances some time ago that traffic to the Congress could be adequately handled); and they are doing all they can from this end to make sure that members are accommodated and that their American offices are alive to the urgency of the situation. Cooks think that the trouble may be in part that it is not always possible at the moment to assure an applicant of a specific sailing date, even though there is no doubt that it will eventually be possible to provide a berth for him about the time he specifies. I understand that additional boats are very likely to be brought into service on the Atlantic route before the summer: but that negotiations on this are not yet quite complete.

Professor Read also found that some American colleagues were dubious about coming over here and making further inroads upon our food supplies. There may also be the very understandable point as to whether it is worth making the trip in view of reports of meagre fare here.

It seems likely that the food situation probably appears worse, in Press reports, than it actually is. We cannot in present circumstances entertain Congress members as lavishly as we should wish. Nevertheless, we have done our best to provide for the comfort and refreshment of members, and think that they may find conditions less rigorous than reports abroad may suggest.

I may mention that, in addition to Government, University and other receptions in London, a large number of civic authorities, industrial firms and other bodies are arranging special local hospitality for the excursion parties.

We greatly hope that despite any present discouraging factors, the number of members from U.S.A. and Canada will approach the number (some 500, including relatives) who provisionally registered in response to the Third Circular. It was largely because of this great response by prospective members from North America that we further extended last autumn our excursion programme; and the success of the Congress will largely depend upon their presence.

Please be assured that on our part we shall spare no effort to satisfy the needs of our guests.

Yours very sincerely,

A. J. BUTLER
*General Secretary*

[1] Personal communication, September, 1947.

VOLUME 56 NUMBER 4

THE JOURNAL OF GEOLOGY

*July 1948*

# INTRODUCTION TO SYMPOSIUM ON PROBLEMS OF MISSISSIPPIAN STRATIGRAPHY AND CORRELATION

J. MARVIN WELLER
University of Chicago

One of the most important results of the work of the Mississippian subcommittee of the National Research Council's Committee on Stratigraphy has been the bringing to light of numerous large problems of stratigraphy and correlation of the Mississippian formations of North America. All these problems have been known to some Mississippian stratigraphers, but few of them have been widely recognized. The existence of some (for example, our sadly deficient knowledge of the Mississippian rocks of the Appalachian region) has probably been unsuspected by most geologists.

The papers of the symposium presented here were solicited by me from among those stratigraphers and paleontologists who seemed best qualified to discuss some of these problems. Selection of problems was necessary in order to hold the symposium within reasonable bounds. In order that the symposium might be of broad interest, the subjects were purposely chosen to encompass a wide geographic range and to include also problems in surface stratigraphy, subsurface correlation, paleogeography, classification, and paleontology. This range of subjects is not exhaustive, however, and numerous other problems of equal, or nearly equal, importance are not touched upon.

The papers of this symposium (except that by Stoyanow) were read before Section E of the American Association for the Advancement of Science in Chicago on December 26, 1947. All are presented here except that by Miller, who feels that further studies are needed before any significant addition can be made to his contribution (in Weller *et al.*, 1948) to the text accompanying the Mississippian Correlation Chart.

The papers of this symposium can be classified roughly in three groups as follows:

1. Those which are concerned principally with important gaps in knowledge and which suggest investigations that are urgently needed. Included in this group are the papers by Byron Cooper, Stockdale, and Williams.

2. Those which are concerned principally with the compilation of data, inadequately presented elsewhere, relevant to certain problems, to which is added more or less new information. Some of

the conclusions reached must be recognized as controversial. These include the papers by Swann and Atherton, Reed, Stoyanow, Miller, Chalmer Cooper, Arnold, and Moore.

3. Those which are more frankly controversial and present viewpoints and conclusions not adequately set forth in other recent publications but not subscribed to by all qualified geologists. These include the papers by Laudon and Selk.

If these papers serve to focus attention on some of the principal problems of Mississippian stratigraphy and correlation in North America and result in the renewal or expansion of investigations designed to solve them, this symposium may be considered to have been successful.

## REFERENCE CITED

Weller, J. M., *et al.* (1948) American Mississippian ammonoid zones. In Correlation of the Mississippian formations of North America: Geol. Soc. America Bull. 59, pp. 91–196.

# STATUS OF MISSISSIPPIAN STRATIGRAPHY IN THE CENTRAL AND NORTHERN APPALACHIAN REGION[1]

BYRON N. COOPER
Virginia Polytechnic Institute, Blacksburg, Virginia

ABSTRACT

Critical study of published information on the Mississippian system of the Appalachian region reveals long and continuing neglect of some of the thickest and most varied sections of the Mississippian in North America.

Abundant faunas and floras in the Appalachian Mississippian have received surprisingly little systematic study and description. These assemblages offer exceptional opportunities for paleontological and biostratigraphical study. Only a beginning has been made toward understanding the local and regional relationships of numerous facies. In many sections of the Appalachian region, the determination of the position, stratigraphic character, and geologic significance of both systemic boundaries awaits detailed study. A large number of correlations and formation identifications, now current, are not only valueless but deceiving because they are based upon inadequate paleontology or in some instances upon no paleontology. The widespread persistence of certain lithologic types has been mistaken for evidence of contemporaneous deposition, with lamentable results.

The keynote of this summary is a strong and urgent plea for renewed interest in Appalachian stratigraphy and for an early beginning to the systematic study of Mississippian fossils.

## INTRODUCTION

The stratigraphic classifications of the Mississippian system in the central and northern Appalachian region embrace three generations of stratigraphic names. "Pocono" and "Mauch Chunk" were introduced by Lesley (1876) as substitutes for the Rogers brothers' "Vespertine" (1844) and "Umbral" (1858), which were the original names applied to the lower and upper parts of the thick Mississippian clastics of northeastern Pennsylvania. W. B. Rogers (MacFarlane, 1879, p. 179) used "Greenbrier limestone" for the limestone below the Mauch Chunk and above the Pocono formation.

The second generation of stratigraphic names, including "Chattanooga," "Grainger," "Fort Payne," "Hinton," "Bangor," "Floyd," "Bluefield," "Princeton," "Bluestone," "Newman," and "Loyalhanna," were introduced during the 1890's and early 1900's, when so many of the classic folios of the United States Geological Survey were prepared. These names were applied to gross lithologic divisions which served as the mapping units of that period.

Third-generation stratigraphic names arose largely from the work of Charles Butts and David B. Reger. Following monographic studies of the Mississippian of Kentucky, Butts (1917, 1922) used Mississippi Valley and Ohio Valley names for some of the divisions of the Appalachian Mississippian. His recognition of the correlatives of the type Mississippian formations was based upon the presence of certain fossils which had been found to be useful guides in the thinner sections prevailing in the Mississippi Valley. In classifying the Mississippian beds from Pennsylvania to Alabama, Butts (1926, 1927, 1932, 1933, 1940) used the names "Warsaw," "St. Louis," "Ste. Genevieve," "Gasper," "Golconda," "Cypress," and "Glen Dean" in a time-stratigraphic sense.

Reger (1926) introduced a long list of stratigraphic names for minor subdivisions of the Mississippian of West Virginia. The names were proposed on the basis of detailed measurement and description of a few stratigraphic sections,

[1] Manuscript received February 18, 1948.

but none of the newly named units was mapped separately. Some of his units in the Greenbrier and Mauch Chunk series have been used in other parts of West Virginia and in some parts of Virginia (fig. 1), but most of the divisions which he recognized have not been traced very far from the sections where they were originally described. No systematic work has been published on the faunas of these units.

Reger (1926) also used some Ohio Valley names for parts of the Appalachian Mississippian, but the names were not used with the support of stratigraphic or paleontological evidence. His recognition of a "Sunbury shale" and "Berea sandstone" is particularly objectionable because their use implies correlations of the most precise character, which are actually based on equivocal evidence. The use of these and other stratigraphic names from the Central Interior region of the United States has misled many geologists into believing that the Mississippian system of the Appalachian region is rather fully understood.

Actually, the stratigraphic nomenclature for the Mississippian of the Appalachian province is most unsatisfactory and is almost wholly inadequate as a framework for understanding the complex variations of the succession. Some elements of the classification are based upon the persistence of certain lithologies whose lower and upper limits are regarded as reliable time boundaries. Facies relationships have been largely ignored. The stratigraphic work upon which the succession was divided into formations was almost entirely of a physical character, and it has been carried on in a near-vacuum of systematic paleontology. Thus many regional correlations are inaccurate.

In order to make rough differentiations of many formations, stratigraphers found it necessary to place heavy reliance upon a few so-called "guide fossils," on the basis of observed ranges in the type Mississippian. The trivial character of some of the paleontological data used in delimiting certain Appalachian formations is exemplified by the determination of 1,500 feet of limestone in the thick Mississippian of Washington County, Virginia, as "Ste. Genevieve limestone" on the basis of the observed range of the columnal plates of *Platycrinites huntsvillae* (Butts, 1940, pp. 369–371).

## GENERAL ASPECTS OF THE MISSISSIPPIAN SYSTEM

In Pennsylvania the Mississippian consists mainly of clastic strata with a total maximum thickness of about 4,000 feet. The lower division, the Pocono sandstone, is well over 1,000 feet thick and is composed primarily of arkosic cross-bedded conglomeratic beds which locally contain plant fossils. The thick Pocono is overlain in northeastern Pennsylvania by the Mauch Chunk red beds. In southwestern Pennsylvania, the Loyalhanna sandy limestone intervenes between the Mauch Chunk red beds and the Pocono. Butts (1924) regarded the Loyalhanna as the much thinned extension of the Ste. Genevieve limestone. The upper Mississippian boundary, as marked by the contact between the Mauch Chunk red beds and the thick, ledge-making Pottsville sandstone and conglomerate, is very sharp. Likewise, the conglomeratic beds in the lower part of the Pocono are rather easily differentiated from the red beds of the Catskill facies.

The generally coarse and nonmarine character of the Pocono of Pennsylvania prevails without much change south-

ward to the latitude of the James River, beyond which the cross-bedded arkosic Pocono strata with plant fossils are gradually supplanted by thin, rusty-weathering sandstones and shales with marine fossils. One thin tongue of Pocono lithology extends as far south as Moccasin Gap, Scott County, Virginia, within a few miles of the Tennessee state line. The rusty-weathering marine beds were named the "Grainger shale" by Keith (1896), but since 1925 the name "Price sandstone" (Campbell, 1925) has been generally used instead of Grainger. Farther southwest toward Cumberland Gap, the Price or Grainger and the underlying Devonian beds with similar lithology are supplanted by a great body of Chattanooga black shale (Hayes, 1891).

The Greenbrier limestones which are so prominently exposed in Virginia and West Virginia are exceedingly thin in Pennsylvania, but in central-western Virginia in the Greendale syncline the Mississippian limestones are more than 4,000 feet thick. Toward Cumberland Gap the Greenbrier limestones become much reduced in thickness and are known as the "Newman limestone" (Campbell, 1893). Part of the Newman is younger than the Greenbrier (fig. 1; Butts, 1940, pp. 405–406).

The Mauch Chunk red beds persist southwestward into West Virginia and Virginia but become variegated and considerably more fossiliferous than they are in Pennsylvania. In the Bluefield area of Virginia–West Virginia, the thick upper part of the Mississippian is composed of ledge- and ridge-making sandstones, red and green variegated fossiliferous shales, thin coal zones with plant fossils, and thin limy members. The lower 1,000–1,800 feet of the succession is the Hinton group of West Virginia (fig. 1) and the Pennington formation of Virginia. The topmost Mississippian formation consists of about 850 feet of beds very similar to the Pennington or Hinton but separated from these lower divisions by a thin persistent conglomeratic sandstone known as the "Princeton formation" (Campbell, 1896). Southward from Bluefield, the Pennington-Princeton-Bluestone succession thins to 150 feet of sandstone at Cumberland Gap. Below this sandstone at Cumberland Gap is a thick limestone, identified by Butts as equivalent to the Glen Dean limestone of the Ohio Valley. The so-called "Glen Dean" at Cumberland Gap is about the same as the Bluefield shale, which underlies the Pennington or Hinton in the Bluefield area.

Figure 1 shows the classification of the Mississippian formations in ten sections between Harrisburg, Pennsylvania, and Cumberland Gap, Virginia-Tennessee-Kentucky. This chart is reproduced almost without change from the Correlation Chart prepared by the Mississippian Subcommittee of the National Research Council (Weller *et al.*, 1948). However, the fossils listed are those particularly relevant to Appalachian stratigraphy. They are not to be considered as absolutely valid guides to the ranges indicated.

The lamentable inadequacy of our knowledge of the Appalachian Mississippian is primarily the result of continuing inactivity in the study of Mississippian faunas of the region. The Mississippian is characterized by facies fully as varied as those prevailing in the Devonian system of New York and in the Appalachian region. The differentiation of the complex facies of the New York Devonian was made possible largely through the accumulation of a wealth of biostratigraphical data from the study of the rich Devonian faunas of that state.

Facies studies of the Appalachian Mississippian have been slowed because of the dearth of systematic work on the fossils.

A comprehensive survey of the stratigraphic problems of the Appalachian Mississippian is beyond the scope of this summary, the primary purpose of which is to stimulate greater interest in the succession than has prevailed during the past. Three fundamental aspects of Mississippian stratigraphy will be discussed in the light of the problems needing special attention.

## SYSTEMIC BOUNDARIES

In central-western Virginia and in adjoining counties of West Virginia the positions of the upper and lower boundaries of the Mississippian system are obscured. In this particular area the position of the systemic boundaries is of great importance because the system therein probably is the fullest Mississippian section on the North American continent.

The lower boundary is within a thick succession of rusty-weathering marine sandstones and shales, the lower part of which is Devonian and the upper part Mississippian. The lower part of the body of sandstone and shale was called "Kimberling shale" and the upper part the "Grainger shale" in some of the early folios of the United States Geological Survey (Campbell, 1896, 1897). More recently, Butts (1932, 1940) has used "Chemung" for the upper part of the Kimberling shale. Since 1925, "Price sandstone" has been used for the beds formerly called "Grainger shale" (Campbell, 1925).

Inasmuch as Butts (1940, pp. 347–350) considered the entire Price formation to be Osagean, he recognized a major hiatus between the Price and the underlying "Chemung." Actually, the Price-"Chemung" succession is transitional in character, with no obvious differentiation of the beds into Mississippian and Devonian components. Butts preferred to draw the base of the Mississippian at the first appearance of *Spiriferina* and *Syringothyris*, but both these genera of brachiopods are now known from Devonian beds. Furthermore, the supposed Osagean age of all the Price formation can no longer be affirmed. Although it is true that the Keokuk age of the upper part of the Price seems to be reliably indicated by a Keokuk fauna containing *Tetracamera subtrigona*, *Orthotetes keokuk*, and *Syringothyris texta*, the lower parts of the Price which contain fossils of early New Providence and Cuyahoga age are probably Kinderhookian (fig. 1). The three species named have been collected from the upper 100 feet of the Price formation near Bluefield, Virginia–West Virginia (Cooper, 1944, pp. 150–151). Two genera of goniatites have been found in the lower Price, *Protocanites lyoni* (Miller, 1936) and *Munsteroceras* (Cooper, 1939, p. 52), both of which occur in the Kinderhookian Chouteau limestone of the type Mississippian.

Recently the writer studied the Devonian-Mississippian succession along New River north of Parrott Station on the Norfolk and Western Railway, Pulaski County, Virginia. Beneath the coarse quartz-pebble conglomerate (Cloyd) at the base of the Price formation and above the highest beds with *Spirifer disjunctus* is a 350–500-foot succession containing fossils of obviously Kinderhookian aspect. A small, rather narrow *Syringothyris*, with a posteriorly deflected ventral cardinal area, and a spiriferoid of the type of *Cyrtospirifer marionensis* are two members of a varied fauna. It is probable that Reger (1926, pp. 505, 520) was deal-

ing with the same fauna and beds when he named the Broadford sandstone from exposures in Smyth County, Virginia. The presence of Kinderhookian beds in the Price and subjacent beds seems established, but the limits of the series are still unknown.

In southwestern Virginia between Cumberland Gap and Bluefield, the Kimberling-Grainger sandstones and shales merge laterally with a thick body of black shale. Where the latter facies predominates, as in western Scott and Lee counties, Virginia, the faunal evidence for drawing the lower systemic boundary is rather meager and equivocal. Although Joel H. Swartz (1926, 1927) has done much to clarify the age relations of various elements of the Chattanooga shale and the probable position of the Devonian-Mississippian boundary, his conclusions should be tested in other areas of the southern Appalachian region.

The upper Mississippian boundary in central-western Virginia and in Mercer County, West Virginia, is somewhere within a thick development of post-Elvira, pre-Pottsville sandstones and shales. The Chester equivalents are over 3,000 feet thick. Very little is known of the faunas of these high Mississippian beds. Near Bluefield, West Virginia, *Sulcatopinna missouriensis*, *Spirifer increbescens*, and *Composita subquadrata* occur in the upper part of the Pennington formation and about 850 feet below the supposed base of the Pennsylvanian Lee formation. These three species collectively delimit the Menard-Kinkaid interval of the Elvira group in the Mississippi Valley. In view of the limited range of these species, the affinities of the faunas occurring in the Bluestone formation need to be investigated. Not only are marine invertebrates plentiful in parts of the Bluestone, but also a well-preserved flora of coal-forming plants is known to occur in the Hunt member.

### FACIES

About eight fundamental facies are recognizable in the Mississippian of the Appalachian region. The general relationships of these facies have not received much study, but certain aspects of the general facies relations are obvious enough to be worthy of mention.

Red beds are prominent in the Mississippian of Pennsylvania. Some of the lower red beds, including the Patton[2] shale, are probably genetically related to the subjacent Devonian Catskill red beds. The Mauch Chunk represents a somewhat different facies, inasmuch as the beds are largely marine. Part of the Mauch Chunk of Pennsylvania is probably equivalent to some of the Greenbrier limestones exposed farther south in West Virginia. The Greenbrier limestones constitute a great sequence of limy beds which divide the Mauch Chunk facies into two parts, the extensions of which reach southward nearly to the Virginia-Tennessee line. The lower of these divisions (fig. 1) is the Maccrady; the upper is the thick Pennington-Princeton-Bluestone succession.

The coarse cross-bedded arkosic beds of the Pocono facies predominate in the lower Mississippian of Pennsylvania and as far south as the James River in Virginia. In the more southeasterly belts of outcrop, tongues of Pocono lithology persist as far south as the Virginia-Tennessee line; locally in New River Valley coal is mined from the Pocono facies, although the Grainger clastic

[2] The name "Patton shale," which appears in fig. 1 and on the Mississippian correlation chart (Weller *et al.*, 1948) is invalid, by reason of prior use of "Patton" for an older shale of Pocono age in western Pennsylvania.

marine facies predominates. In Smyth, Tazewell, and Washington counties, the Price-Grainger lithology prevails to the almost complete exclusion of the Pocono facies. Still farther southwestward, in the Cumberland Gap district, the Grainger lithology is supplanted by the Chattanooga black shale. The complementary relationship of the Grainger-Kimberling marine clastics, as developed in the Bluefield area, to the Chattanooga shale in the Cumberland Gap area is one of the most obvious facies changes in the Paleozoic sequence of the southern Appalachian region (fig. 2).

The Greenbrier limestone magnafacies, with its maximum thickness in Virginia and West Virginia, is supplanted to the northeast by the Mauch Chunk and the Pocono. The time represented by the type Greenbrier is probably recorded in the Pocono and the Mauch Chunk of Pennsylvania (fig. 2). The Greenbrier limestone facies grades into the clastic facies above and below it not only by intertonguing but also by almost imperceptible lithologic transitions. Indeed, the entire Greenbrier facies is enclosed by a fringe of hybrid lithologies. The impure limestone above the Maccrady red beds and below the main body of Greenbrier limestone in the Greendale syncline of Washington County, Virginia, is a Maccrady-Greenbrier hybrid parvafacies. The Bluefield formation bears the same relationship to the Green-

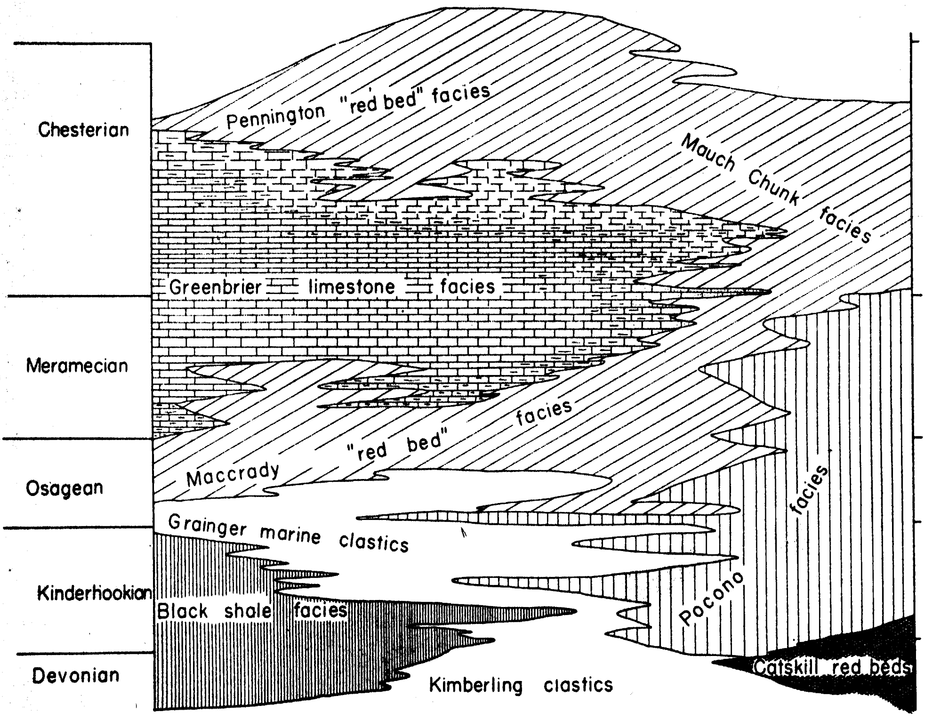



FIG. 2.—Idealized relations between facies in the Mississippian system between Harrisburg, Pennsylvania (*right*) and Cumberland Gap, Virginia (*left*).

brier and Pennington-Bluestone beds. The Loyalhanna limestone is a hybrid facies related to the Pocono and Greenbrier (fig. 2). The Cove Creek and Glen Dean limestones of Virginia are also hybrid facies related to the Pennington-Bluestone division.

The variations within each of these facies are very complicated. For example, the writer in the course of routine field investigations on the industrial possibilities of the Greenbrier limestones of southwestern Virginia came to recognize fourteen distinctive types of limestone, which are repeated at different stratigraphic levels. These parvafacies are not as difficult to separate as are those of the Grainger-Price-Pocono beds, which are the product of both marine and nonmarine deposition.

### MISSISSIPPIAN SECTION IN THE GREENDALE SYNCLINE OF VIRGINIA

A summary of Mississippian stratigraphy should include mention of the amazing section which occurs in the Greendale syncline of Virginia (fig. 3). In many respects this succession is without an equal elsewhere in North America. More than half the 7,000-foot thickness is fossiliferous limestone. The general

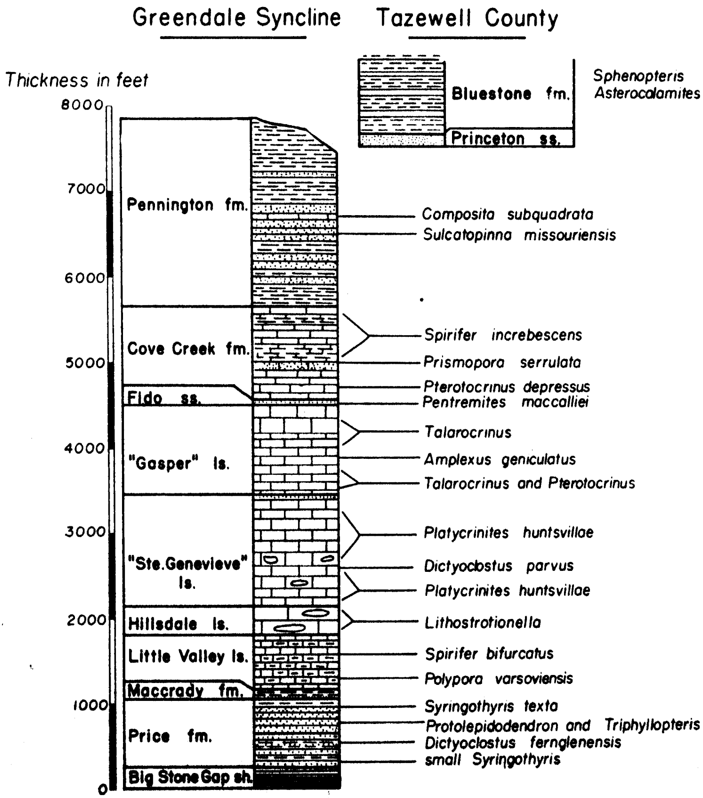



FIG. 3.—Mississippian formations in central-western Virginia

features of the section have been described by Butts (1940, pp. 337–340) and Averitt (1941); but most of the formations that they recognized are gross lithologic units.

At the base of the section is a post-Chemung black shale with *Lingula*, *Leiorhynchus*, and *Sporangites*. Succeeding this 60-foot unit is about 800 feet of Price sandstone with *Spirifer marionensis* and *Syringothyris sp.* in the lower part. Beds somewhat higher contain *Euphemites galericulatus* and *Dictyoclostus fernglenensis*, and the upper 200 feet contain *Pseudosyrinx* and *Orthotetes keokuk*. Above the Price are the Maccrady red beds, here very thin and containing *Polypora varsoviensis* and other Warsaw bryozoans. The overlying Little Valley limestone is about 700 feet thick and contains the same bryozoan fauna as is present in the subjacent Maccrady red beds and other Warsaw-Salem fossils, including *Camarotoechia mutata* and *Spirifer bifurcatus*.

The next unit is the Hillsdale limestone, with the ubiquitous guide *Lithostrotionella* indicating general equivalency to the St. Louis limestone of the Mississippi Valley. Above the 300-foot St. Louis equivalent are no less than 1,500 feet of limestone, which seem to be older than the Renault limestone (fig. 1) of the type Mississippian. The 1,500-foot division contains from top to bottom a profusion of stem plates belonging to *Platycrinites* of the type of *P. huntsvillae*, on which Butts (1940, pp. 372–374) relied heavily for identification of the Ste. Genevieve limestone. Other supposed Ste. Genevieve guides, including *Dictyoclostus parvus* and *Pentremites princetonensis*, also occur, but the precise range of the Ste. Genevieve fauna has not been determined.

The Ste. Genevieve limestone equivalent is overlain by 1,000 feet of limestone containing *Pterotocrinus* and *Talarocrinus* and also a coral, possibly the same as the Renault guide, *Amplexus geniculatus*. This "Gasper" limestone is overlain, in turn, by a thin ferruginous sandstone, the Fido, containing the giant blastoid, *Pentremites maccalliei*. This species is very closely related to the *Pentremites obesus* of the type Mississippian Golconda formation. This sandstone is followed by the 1,000-foot Cove Creek limestone, which abounds in Glen Dean fossils and carries the distinctive genus *Prismopora*—a guide to the Golconda-Vienna interval of the Mississippi Valley region. The Cove Creek limestone is overlain by Pennington-type sandstones and shales, with late Chester fossils in the middle portion of the 1,000–1,600 feet of the Pennington (fig. 3). If the Princeton and Bluestone formations of adjoining Tazewell County, Virginia, are younger than the Pennington beds of the Greendale syncline, then approximately 900 feet of beds must be added to complete a full composite section of the Mississippian system of southwestern Virginia.

In comparing the Mississippian section of the Greendale syncline with the type Mississippian, it is barely possible that a fuller depositional record is being compared to a less nearly complete, though better-known, section. At least, one can expect to find the faunas of the type Mississippian deployed through greater stratigraphic thicknesses in the central Appalachian region than in any other part of the United States. The attractiveness of the Mississippian succession of the Greendale syncline as a field for biostratigraphic and paleontological studies is enhanced by the almost continuous exposures along several main highways and by extensive glady areas

wherein well-preserved fossils are abundant.

## SUMMARY

Although the Appalachian region received considerable study in the early days of reconnaissance mapping by the United States Geological Survey, the region has received relatively little detailed study. Most of the Appalachian faunas are still undescribed or only poorly known, and few attempts have been made to work out regional and local facies relationships in any of the Paleozoic systems. This unsatisfactory state of knowledge applies particularly to the Mississippian system, which contains many essentially untouched fields for stratigraphic and paleontological study. However, the same types of challenging problems pertain to the other Paleozoic systems of the Appalachian region. The progress of modern stratigraphic work in this region will continue to be exceedingly slow as long as paleontological studies are no more active than they have been in the past.

## REFERENCES CITED

AVERITT, PAUL (1941) The Early Grove gas field, Scott and Washington counties, Virginia: Virginia Geol. Survey Bull. 56, pp. 7–29.

BUTTS, CHARLES (1904) Description of the Kittanning quadrangle, Pennsylvania: U.S. Geol. Survey Geol. Atlas, Folio 115.

——— (1917) Descriptions and correlation of the Mississippian formations of western Kentucky, Kentucky Geol. Survey.

——— (1922) The Mississippian series of eastern Kentucky: Kentucky Geol. Survey, ser. 6, vol. 7.

——— (1924) The Loyalhanna limestone of southwestern Pennsylvania especially with regard to its age and correlation: Am. Jour. Sci., 5th ser., vol. 8, pp. 249–257.

——— (1926) The Paleozoic rocks, in Geology of Alabama: Alabama Geol. Survey Special Rept. 14, pp. 162–208.

——— (1927) Oil and gas possibilities at Early Grove, Scott County, Virginia: Virginia Geol. Survey Bull. 28.

——— (1932) Southern Appalachian region: 16th Internat. Geol. Cong. Guidebook 3, pp. 22–24.

——— (1933) Geologic map of the Appalachian Valley of Virginia with explanatory text: Virginia Geol. Survey Bull. 42, pp. 36–46.

——— (1940) Geology of the Appalachian Valley in Virginia: Virginia Geol. Survey Bull. 52.

CAMPBELL, M. R. (1893) Geology of the Big Stone Gap coal field: U.S. Geol. Survey Bull. 111, pp. 26, 38.

——— (1896) Description of the Pocohantas sheet, Virginia: U.S. Geol. Survey Geol. Atlas, Folio 26.

——— (1897) Description of the Tazewell quadrangle, Virginia: *ibid.*, Folio 44.

——— (1925) The valley coal fields of Virginia: Virginia Geol. Survey Bull. 25, pp. 23–28.

COOPER, B. N. (1939) Geology of the Draper Mountain area, Virginia: Virginia Geol. Survey Bull. 55, pp. 47–54.

——— (1944) Geology and mineral resources of the Burkes Garden quadrangle, Virginia: *ibid.*, Bull. 60, pp. 145–187.

HAYES, C. W. (1891) The overthrust faults of the southern Appalachians: Geol. Soc. America Bull. 2, pp. 141–154.

KEITH, ARTHUR (1896) Description of the Morristown sheet, Tennessee: U.S. Geol. Survey Geol. Atlas, Folio 27.

LESLEY, J. P. (1876) Historical sketch of geological explorations in Pennsylvania and other states; with an appendix containing the annual reports of the state geologist to the Board of Commissioners: Pennsylvania Geol. Survey 2d Ann. Rept., pp. 221–227.

MACFARLANE, JAMES (1879) An American geological railway guide giving the geological formation at every station, New York.

MILLER, A. K. (1936) A Mississippian goniatite from Virginia: Jour. Paleontology, vol. 10, pp. 69–72.

REGER, D. B. (1926) Mercer, Monroe, and Summers counties: West Virginia Geol. Survey, pp. 316–520.

ROGERS, H. D. and W. B. (1844) Address on the recent progress of geological research in the United States: Am. Jour. Sci., vol. 47, pp. 137–160.

——— — ——— (1858) The geology of Pennsylvania, vol. 1, pp. 108, 142–144.

SWARTZ, J. H. (1926) The age of the Big Stone Gap shale of southwestern Virginia: Am. Jour. Sci., 5th ser., vol. 12, pp. 512–531.

——— (1927) The Chattanoogan age of the Big Stone Gap shale: *ibid.*, 5th ser., vol. 14, pp. 485–499.

WELLER, J. M., *et al.* (1948) Correlation of Mississippian formations of North America: Geol. Soc. America Bull. 59, pp. 91–196.

# SOME PROBLEMS IN MISSISSIPPIAN STRATIGRAPHY OF THE SOUTHERN APPALACHIANS[1]

PARIS B. STOCKDALE
University of Tennessee

ABSTRACT

Mississippian stratigraphic problems of the southern Appalachians pertain to: (1) refinement in stratigraphic subdivision; (2) establishment of lateral relationships from one place to another, especially between disconnected areas and areas of independent previous study; (3) correlation of the southern Appalachian strata with the better-known units of the standard section; (4) establishment of the manner and time of origin of some of the units; and (5) establishment of the Mississippian-Pennsylvanian boundary. Examples of major problems which need further study are discussed. Solution of many of the stratigraphic problems must await careful and detailed field studies and, particularly in several instances, a consideration of the faunas.

It would be presumptuous for the writer to pose as an authority on the problems of Mississippian stratigraphy in the southern Appalachians, because his own researches have been confined to the Mississippian of the North-central Interior. Nevertheless, summer field courses on the Cumberland Plateau in eastern Tennessee, reconnaissance field excursions in the southern Appalachians, and perusal of the literature have acquainted him with some of the problems. However, only a few of the outstanding problems will be discussed. For the sake of brevity, no attempt is made to summarize the works of previous writers or to cite the literature in full. The author's responsibility, as he sees it, is to give a general summary of the problems of Mississippian stratigraphy in the southern Appalachian area.

Foremost among present-day needs is a refinement in stratigraphic subdivision, together with closer correlation. Many units, now recognized throughout much of the area, are thick and highly generalized. A good example is the Bangor (restricted) limestone, a generalized group of limestone beds of Chester age, 100–700 feet thick, which should be subdivided and related to the known divisions of the Chester series of the Mississippi Valley. The Chester of eastern Kentucky should be traced southward; and its fauna needs careful collecting and study. The Bangor group, as now defined, is a "restricted" survival of the original Bangor of the early literature, which included all Mississippian strata above the Fort Payne chert. Later workers identified the St. Louis, Ste. Genevieve, and Gasper limestones in the lower part of the group and separated the Pennington shale from the topmost portion. The undifferentiated remainder is all that is now called "Bangor." A calcareous, ferruginous sandstone, locally present between the top of the Gasper and the base of the restricted Bangor, has been called "Hartselle." This sandstone is generally believed to be the age-equivalent of the Hardingsburg sandstone, although agreement on this point is not unanimous. According to Butts (1926), the Bangor "restricted" should be correlated with the combined Glen Dean, Tar Springs, Vienna, and Waltersburg beds of the standard section. These beds have not been differentiated in the southern Appalachians.

Another Mississippian problem in the

[1] Manuscript received February 7, 1948.

southern Appalachian area is the *lateral* relationship between the rock of a given outcrop belt and that in another belt not too far away. The lithological dissimilarity is, of course, a facies problem. The so-called "Floyd shale" affords an example. About midway between the eastern and western margins of the Valley and Ridge Province there are several outliers of this shale, up to 2,000 feet thick, which trend northeast-southwest across southern Tennessee, northwestern Georgia, and Alabama. This shale overlies the Fort Payne chert. The name "Floyd" has been abandoned in some localities where correlations with the standard section have been suggested. However, there remains considerable uncertainty as to the relationships in other localities. For example, in the Ringold and Chattanooga quadrangles, the Floyd shale is quite prominent along one belt, whereas a few miles to the west, shale is missing, and only a limestone facies is present. Butts (1926), in his excellent studies in Alabama, concluded that the "Floyd corresponds to the formation . . . from the base of the Gasper to the top of the Bangor, and probably extends as low as the base of the Ste. Genevieve." The age relationships between the rocks of these two belts are not clear and call for detailed field and faunal studies.

Toward the northern end of the midsection of the Valley and Ridge Province in Tennessee is another thick shale, called the "Grainger," which lies above black, fissile shale and below so-called "Fort Payne chert." Like the Floyd shale, it, too, is absent in the sections at the western edge of the province a few miles away. Considerable uncertainty remains as to the stratigraphic relationship of the Grainger to the typical Chattanooga black shale farther southwest, the Olinger member of Swartz farther northeast, and other Mississippian strata at the plateau escarpment a short distance farther west, where the Grainger is absent. On the east side of the Valley and Ridge Province, lying directly against the boundary of the Older Appalachians, at the base of Chilhowee Mountain in the Knoxville quadrangle, is a little-known belt of shale and sandstone which also was called "Grainger" by Keith (1895). These beds, 1,000 feet thick, present one of the important, though smaller, problems in southern Appalachian Mississippian stratigraphy, that has received little study.

Even where attempts have been made to identify and to separate specific formations within generalized units, such as the Bangor, there is still considerable uncertainty in the placing of exact boundaries in the field. For example, at many exposures along the Cumberland Plateau or along the sides of the Sequatchie Valley, the exact contact between the St. Louis and the Ste. Genevieve has not been agreed upon; and the boundary between the latter and the Gasper is somewhat arbitrary and uncertain. Again, there is seen the need for a more thorough collecting and study of fossils.

The "black-shale problem" is a major one in the southern Appalachians, as it is elsewhere. The possible subdivision of the Chattanooga shale into members, the manner and time of origin of the sediments, and the relationships with the New Albany and Ohio black shales to the north and northeast in Kentucky, Virginia, Ohio, and Indiana constitute problems which are too large for full discussion in this paper. Many stratigraphers are, of course, familiar with the general Devonian-Mississippian age controversy of these black shales. After working in the Highland Rim of Tennessee, Klepser contends that the black-shale facies,

from southern Kentucky southward, extends into a much higher portion of the column, and he concludes that " . . . as the Chattanooga and overlying Maury everywhere grade upward into the overlying units of successively younger age southward, it must be concluded that the Chattanooga is a time-transgressing unit ranging from Kinderhook to Keokuk in age." Following an early suggestion of Grabau, he contends, in other words, that the shales represent the basal shore phase of a sea advancing slowly southward upon low-lying swampy land and that their northern time-equivalents are the normal, off-shore sediments of the New Providence and Fort Payne formations. Klepser presented his views in a doctoral dissertation (1937) and in a paper presented before the Tennessee Academy of Science in November, 1946. The absence of the New Providence shales in eastern Tennessee certainly calls for an explanation. In Kentucky, Ohio, and Indiana, the New Providence shale is a progressively northward-thickening wedge, which separates the black shale beneath from the younger Keokuk siltstones above. The microfaunas, especially the conodonts, have led various workers to suggest correlations between the subdivisions of the Chattanooga black shale in the southern Appalachians and the black shales farther north. The recent studies of W. H. Hass bearing on this question were summarized in the published abstracts of papers listed for the December, 1947, meetings of the Geological Society of America. In the abstract of his paper, entitled "The Chattanooga Shale Type Area," Hass stated:

> At the Apison locality, the upper black shale member contains lower Mississippian conodonts and is correlated with the Sunbury shale of Ohio. The lower black shale member contains conodonts that correlate it with the Huron shale of Ohio, a formation that the U.S. Geological Survey classifies as Upper Devonian. The middle gray shale member contains Huron conodonts, but its age is equivocal as J. H. Swartz has reported macrofossils from it which he considered to be of early Mississippian age. . . . The presence of Huron conodonts in the lower black shale member of the Chattanooga disproves the thesis, held by some workers, that, as a unit, the Chattanooga shale is younger than the black shale sequence of the North-Central States.

This line of reasoning, shared also by others, that identical or closely similar assemblages of microörganisms in deposits at widely separated localities prove the contemporaneous age of the sediments should, in the author's opinion, be carefully weighed against the thesis that a faunal assemblage might have persisted under a given environmental condition across a wide span of time. If all or but a part of the black shales in question, which stretch across hundreds of miles of territory, are a "magnafacies," or "phase," deposited in an environment which shifted slowly southward with the passage of time, is it not conceivable that a given assemblage of microörganisms favorable to a given set of paleoecological circumstances might have remained unchanged throughout a considerable span of time and might now be found as a fossil assemblage coextensive with the given lithologic, time-transgressing unit? Such a paleoecological, or facies-fauna, concept may resolve some of the stratigraphic perplexities which arise from field studies. Certainly, it is one that was impressed upon the writer by his own field studies of the Lower Mississippian formations of southern Indiana and the East-central Interior (1931, 1939).

The author's detailed studies of the Lower Mississippian strata of the East-central Interior, in Indiana, Kentucky,

and parts of adjoining southern Ohio, especially the great thickness of Osage shales and siltstones, leads him to ask certain questions about these rocks in the southern Appalachians: (1) Why are they missing to the south, and what happens to them in the concealed cover of the Appalachian Plateau? (2) Are the calcareous rocks of the Fort Payne formation of eastern Tennessee and northern Alabama and Georgia the time equivalents of the clastic sediments of the Borden group of Indiana, the New Providence, Brodhead, and Muldraugh of Kentucky, and the Cuyahoga and Logan of Ohio, or are these strata represented by a profound unconformity? (3) Where did these sediments come from?

Nearly everywhere in the southern Appalachians, the Fort Payne beds lie evenly upon the thin and persistent Maury shale with its phosphatic nodules, which, in turn, overlies the Chattanooga black shale. Above the Fort Payne and between it and the St. Louis (Meramec) limestone, in eastern Tennessee, occur some drab-colored siltstones with worm-trails and *Taonurus* markings. Because these strata weather readily, they are generally concealed and are little known. More recent workers have referred them to the Warsaw. Their relationships to the clastics to the north are not understood and need study. Certainly, these siltstones are lithologically similar to the Keokuk (Osage) siltstones of Kentucky. This brings up the classic Mississippian problem—the boundary between the Osage and the Meramec. In tracing the Mississippian rocks from Indiana southward and eastward across the Kentucky outcrop belt, the writer (1939) was impressed by the conspicuous unconformity at the base of the St. Louis, which transgresses underlying beds and leads to the gradual eastward disappearance of Warsaw strata. This relationship not only indicates the prominence of the break between the Osage and the Meramec groups but gives additional support to other field evidence that the Warsaw strata are more closely allied to the Osage than to the Meramec.

Nearly everywhere in the southern Appalachians, the Fort Payne chert lies in even beds upon the Chattanooga-Maury sequence. Recently, however, the writer found in the White Oak Mountain area an exposure of a large bioherm, composed of myriads of crinoids, projecting prominently upward into the Fort Payne. A careful study of this bioherm and search for others may contribute to the solution of the Fort Payne–Chattanooga age relationships. Such bioherms, which are well known in the Mississippian of the East-central Interior, have not before been reported from the southern Appalachians.

In a summary of the problems of the Mississippian stratigraphy, the problem of the Mississippian-Pennsylvanian boundary cannot be omitted. Whereas, throughout the greater part of the North American interior, a marked unconformity separates the two systems, such a physical break is missing, or is certainly not clear, throughout much of the southern Appalachians. In Alabama the Parkwood formation, 2,000 feet of shale and arkosic sandstone, closely resembles the overlying rocks of unequivocal Pennsylvanian age. Its lower part is believed to grade laterally into the Pennington shale in eastern Tennessee. In 1926 Butts wrote: "The Parkwood bridges the gap or unconformity that elsewhere in the Appalachian region intervenes between the Mississippian series and Pennsylvanian series." In eastern Tennessee the Pennington shale, which is uppermost Chester in age, is overlain by the Gizzard

formation of Pottsville age. Many fine exposures have been studied along the Cumberland escarpment at the eastern margin of the Appalachian Plateau, where a dearth of fossils and a gradual gradation upward from the clayey shales of the typical Pennington into the sandy shales of the Pottsville makes it impossible to indicate other than an arbitrary systemic boundary.

In this paper, the author has confined himself to the limitations of the title—*problems* in Mississippian stratigraphy of the southern Appalachians. Many of the solutions and answers rest with the future.

## REFERENCES CITED

BUTTS, CHARLES (1926) Geology of Alabama: the Paleozoic rocks: Alabama Geol. Survey Special Rept. 14, pp. 41–230.

HASS, W. H. (1947) The Chattanooga shale type area (abstr.): Geol. Soc. America Bull. 58, p. 1189.

KEITH, ARTHUR (1895) Description of the Knoxville sheet: U.S. Geol. Survey Geol. Atlas, Knoxville Folio.

KLEPSER, H. J. (1937) The Lower Mississippian rocks of the eastern Highland Rim: Abstracts of Doctor's Dissertations, no. 24, Ohio State University.

STOCKDALE, P. B. (1931) The Borden (Knobstone) rocks of southern Indiana: Indiana Dept. Cons., Div. Geology Pub. 98.

——— (1939) Lower Mississippian rocks of the east-central interior: Geol. Soc. America Special Paper 22.

# SUBSURFACE CORRELATIONS OF LOWER CHESTER STRATA OF THE EASTERN INTERIOR BASIN[1]

DAVID H. SWANN AND ELWOOD ATHERTON
Illinois Geological Survey, Urbana, Illinois

ABSTRACT

Direct comparison of logs of closely spaced wells and their assembly into cross sections appear to demonstrate a number of inconsistencies in the nomenclature that is currently applied to the lower part of the Chester series and to the Ste. Genevieve formation in different parts of the Eastern Interior Basin. This evidence indicates that: (1) the important oil-producing "Aux Vases" sand of the central basin area is equivalent to the outcropping Rosiclare sandstone member of the Ste. Genevieve formation; (2) the "Benoist" (Yankeetown), Bethel, and Sample sandstones are successively younger rather than correlative units; (3) the Cypress sandstone of Indiana and west-central Kentucky is younger than the Cypress of Illinois; (4) certain continuous limestone units can be traced between the various areas (the Beech Creek or "Barlow" limestone lies above the Illinois Cypress but beneath the Indiana Cypress); (5) the Downeys Bluff limestone, for a long time considered a part of the Renault formation because it lies below the Bethel sandstone, is in reality the basal Paint Creek limestone above the Yankeetown and is correlated with the upper portion of the Paoli limestone of Indiana; (6) the Levias limestone and part of the Shetlerville (Renault) together form a continuous limestone sequence which overlies both the "Aux Vases" of the basin and the Rosiclare sandstone of the outcrop area in southeastern Illinois.

## INTRODUCTION

The intensive drilling program that began in 1937 when oil was discovered in the deeper part of the Eastern Interior Basin has produced a wealth of stratigraphic data on the Mississippian formations above the St. Louis limestone. Almost all locations within the Illinois Basin proper (the deepest part of the Eastern Interior Basin) are within 2 miles of wells or tests for which good stratigraphic records are available. Thus a re-examination of the long-range correlations made at the beginning of the basin oil development appears in order.

The fund of information on the variation in stratigraphy indicated by subsurface records provides material not only for subsurface correlation but also for a new attack on the perplexing problems of outcrop correlation around the border of the basin area.

This paper records the stratigraphic columns in current use in different parts of the Eastern Interior Basin for the Mississippian strata between the St. Louis formation of the Meramec group and the Hardinsburg sandstone of the Homberg (middle Chester) group. Correlation between the several areas is indicated by a chart (fig. 1). The chart is substantiated by the cross sections (figs. 3–6) whose locations are shown on the index map (fig. 2). The conclusions are similar to those of Dana and Scobey (1941) but are here presented in greater detail.

Although this report indicates the need for a revision of the standard classification of the lower part of the Chester series, this revision is not attempted here. It is postponed, awaiting clarification of several problems involving the correlation of formations defined in the southwestern Illinois and Missouri outcrop area with the units which can be traced throughout the rest of the basin.

## CORRELATION CHART

The left-hand column of the correlation chart (fig. 1) has been slightly modified since the Chester rocks in south-

[1] Published by permission of the Chief, Illinois Geological Survey, Urbana, Illinois. Manuscript received May 20, 1948.

western Illinois were described by S. Weller (1913, p. 120); its present form is that used by Cooper (1941, p. 7). The formation boundaries as originally given by Weller are in current use and appear to be applied consistently throughout the outcrop area, which extends along the Mississippi River for 85 miles below St. Louis. This area is west of the limits of the index map (fig. 2).

The second column gives the usage applied by members of the Illinois Basin Oil Scouts Association in the Louden, Salem, Boyd, and Woodlawn oil fields of south-central Illinois and in the area west of these fields. Boyd (fig. 2) is near the eastern edge of the area in which this nomenclature is consistently used.

The third column shows the nomenclature applied by the geologists of oil companies operating in the lower portion of the drainage basin of the Wabash River in Illinois and Indiana. Maunie and New Haven, Illinois, are in this area (fig. 2).

The fourth column of figure 1 shows the names applied in the fluorspar area of Hardin County, Illinois, and adjacent parts of Illinois and Kentucky by S. Weller (1920*b*, pl. 1), as modified by J. M. Weller and Sutton (1940, p. 766), and Atherton (1948, p. 129). Golconda, Illinois, is in this region (fig. 2).

The names currently applied in oil fields in Henderson and Daviess counties, Kentucky, are indicated in the fifth column of figure 1. Poole oil field (fig. 2) is near the heart of this area.

The sixth column is the outcrop section in west-central Kentucky given by Stouder (1941, p. 25), and the last column is the outcrop section in Indiana given by Malott (1931, p. 222 and 1946, pp. 322–326). Leavenworth, Indiana (fig.

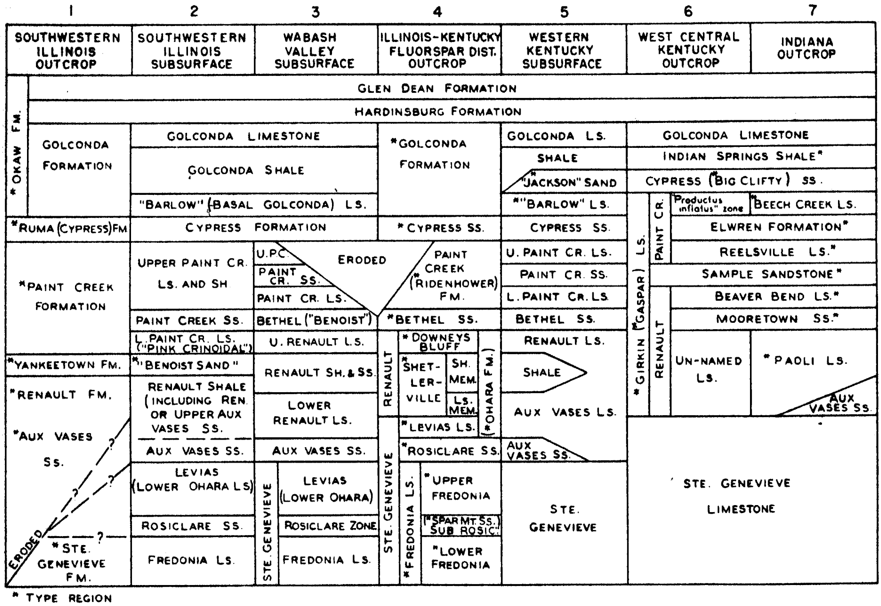



FIG. 1.—Correlation of middle Chester, lower Chester, and Genevieve strata in Eastern Interior Basin

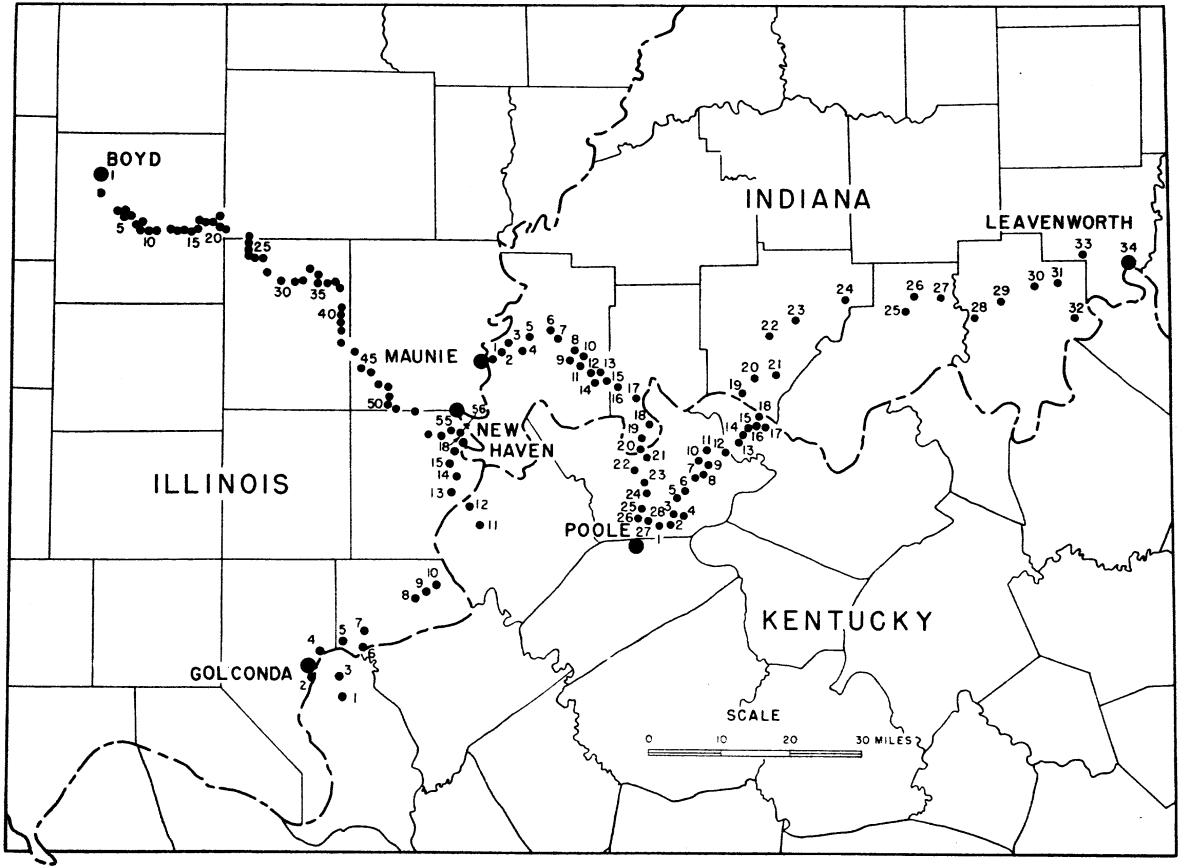



FIG. 2.—Index map of part of Eastern Interior Basin, showing wells and outcrops used in cross sections

2), lies in the outcrop belt of southern Indiana and west-central Kentucky which is described by Malott and Stouder.

The correlations between the first and second columns of figure 1—the southwestern Illinois outcrop and subsurface areas—are not represented by a cross section. As indicated by Tippie (1943), these correlations are commonly accepted in this region. They are further discussed in the stratigraphic section of this report under "Beech Creek limestone," "Downeys Bluff limestone," and "'Benoist' (Yankeetown) sandstone."

## CROSS SECTIONS

The geologic cross sections (figs. 3–6) pass through wells nowhere more than 11 miles apart and averaging less than 2 miles apart. Most sections were compiled from electric logs. Although only a few major lithologic types can be indicated on the figures, minor distinctions in lithology were used throughout the study to guard against errors in correlation between adjacent wells. The original cross sections were drawn on a vertical scale of 20 feet to 1 inch. Twelve to fourteen different physical types of rock were classified by their electrical characteristics, and the electric-log interpretations were checked by sample studies wherever there appeared to be any possibility of miscorrelation. All cross sections have a vertical exaggeration of 211.

Figures 3, 4, and 5 have been prepared by using the base of the Downeys Bluff limestone as the datum plane. Figure 3 substantiates the correlations indicated between the second and third columns of figure 1 and shows that the "Benoist" (Yankeetown) and the Bethel sandstones are not continuous. Figure 4 shows that the terminology used for the lower part of the stratigraphic section in the Wabash Valley oil fields is a full sedimentary cycle lower than that applied to the same beds in the outcrop area of southeastern Illinois. Figure 5 carries the correlation from the Wabash Valley to the heart of the western Kentucky oil fields across the barrier of continuous Bethel-Sample-Cypress sandstone deposition.

Figure 6 has been prepared with the Beech Creek or "Barlow" limestone as a datum plane. This has been done because the Downeys Bluff, which is readily distinguishable throughout Illinois and the adjacent portions of Indiana and Kentucky, becomes difficult to differentiate from the other beds of the Paoli limestone as the Indiana-Kentucky outcrop belt is approached. This cross section extends from the western Kentucky oil-field district to the Indiana outcrop belt, substantiating the correlations indicated between the fifth column and the last two columns of figure 1. The Beech Creek or "Barlow" limestone is shown to underlie the Big Clifty or Cypress sandstone of the Indiana section and to overlie the Illinois Cypress. The Bethel sandstone of western Kentucky is equivalent to the Mooretown sandstone rather than to the Sample sandstone of the outcrop belt. Figure 6 shows the entire Chester series rather than just the lower portion, as seen in the preceding cross sections. It indicates clearly the character of the pre-Pennsylvanian surface and the regular thinning of most Chester units toward the eastern outcrop belt. The correlations of the upper part of the Chester series demonstrated here agree with those given by Malott (1931, p. 222). The Indiana names for this part of the section have been virtually abandoned in favor of their western correlatives.

The logs of fluorspar tests[3] in Living-

[3] H. H. Cronk, superintendent of the Rosiclare Lead and Fluorspar Mining Company, and P. L. Richards, superintendent of the Inland Steel Com-

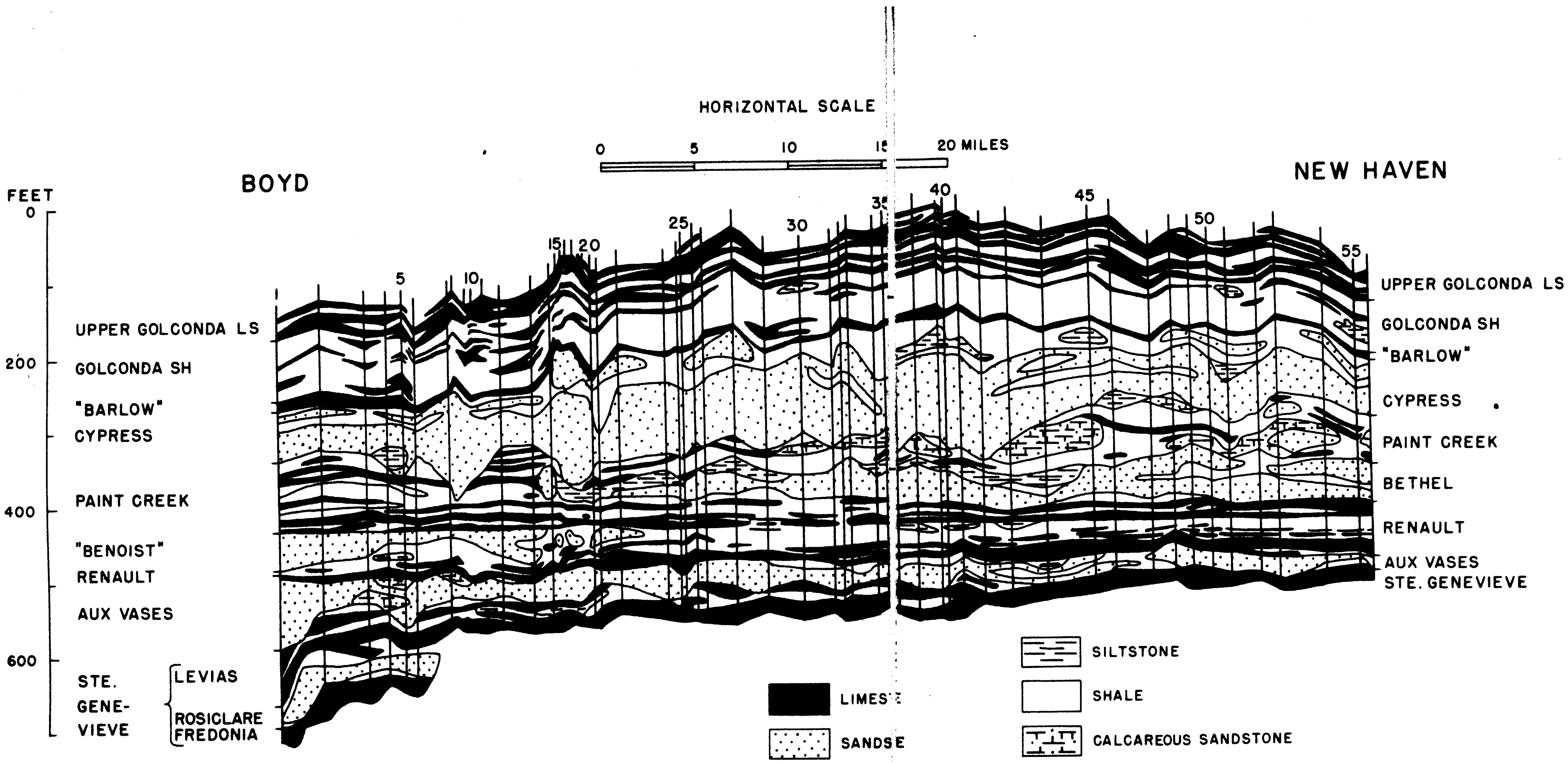



FIG. 3.—Cross section of lower Chester strata between oil fields of southwestern Illinois near Boyd and Wabash Valley oilds near New Haven, Illinois (see fig. 1, cols. 2 and 3). Well lines omitted in congested portions

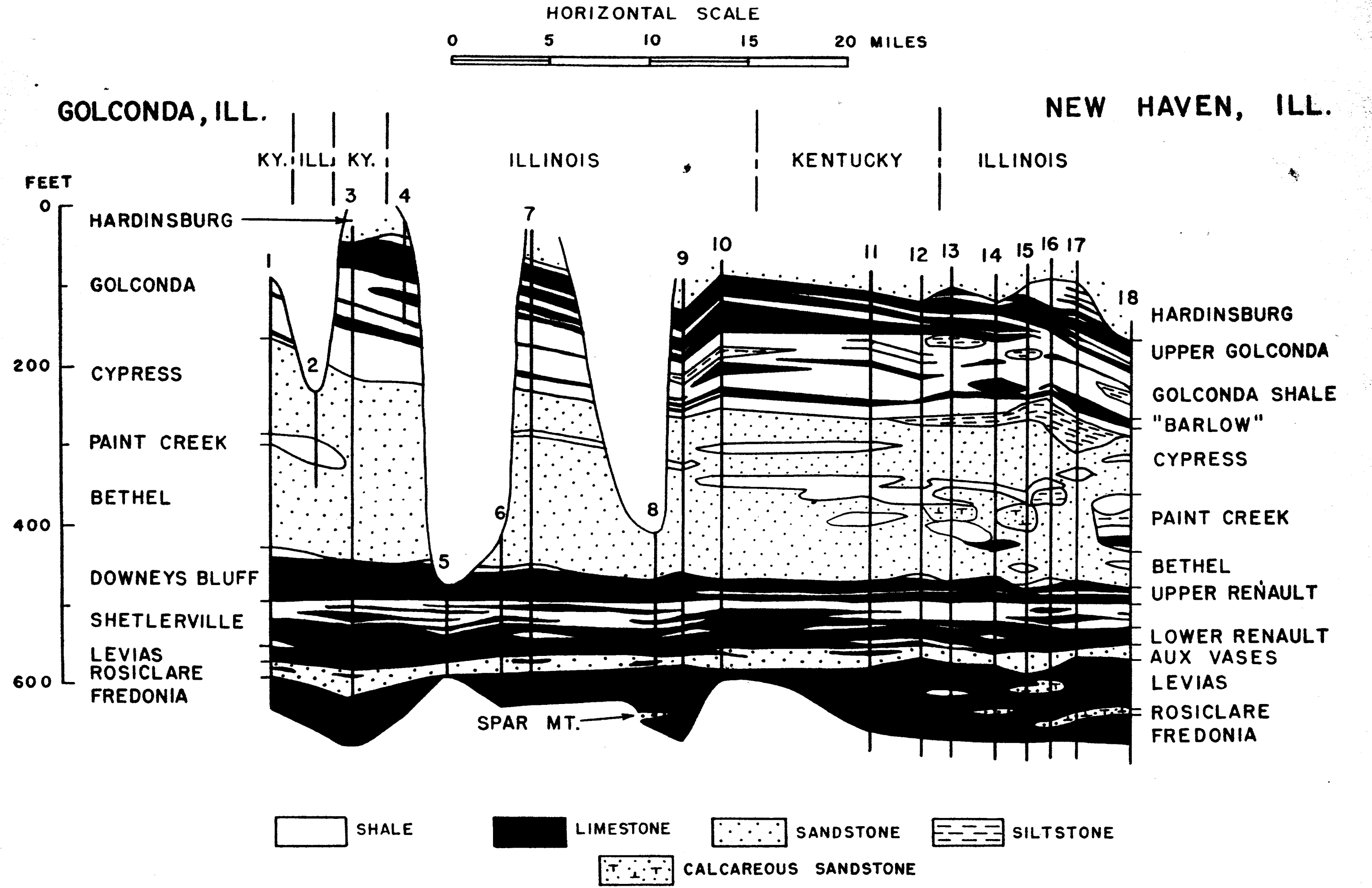



FIG. 4.—Cross section of lower Chester strata between outcrops in the Illinois-Kentucky fluorspar district near Golconda and the Wabash Valley oil fields near New Haven, Illinois (see fig. 1, cols. 4 and 3).

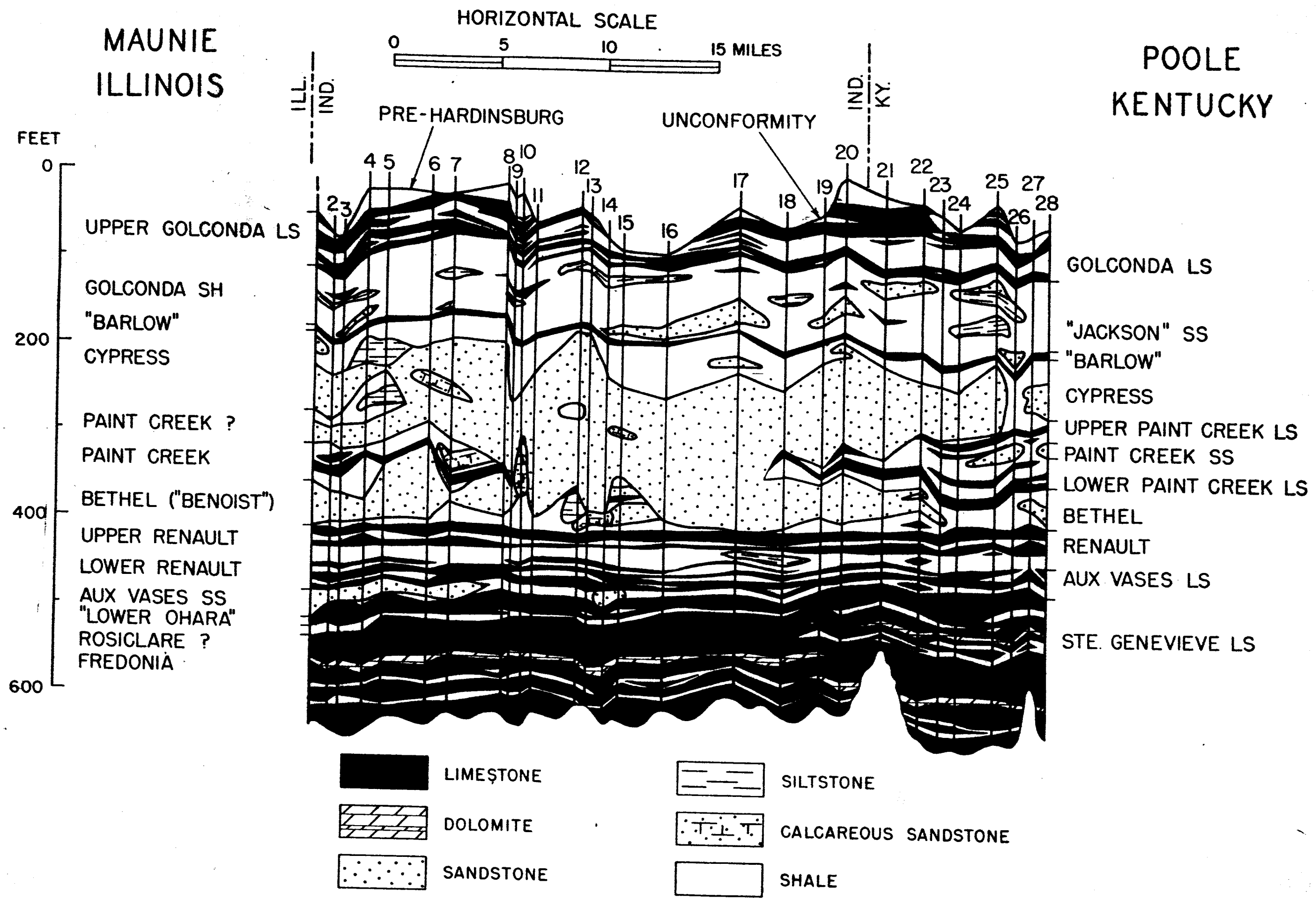



FIG. 5.—Cross section of lower Chester strata between the Wabash Valley oil fields near Maunie, Illinois, and the western Kentucky oil fields near Poole (see fig. 1, cols. 3 and 5).

ston County, Kentucky, and Hardin County, Illinois, used in figure 4 are based on descriptions by Gill Montgomery of the Minerva Oil Company and by several members of the Illinois Geological Survey. Section 2 of figure 4 is that given by S. Weller (1920, p. 168); section 4 of this figure was measured by L. E. Workman and published by Atherton (1948, p. 128); sections 5 and 6 have been described in several publications on the stratigraphy of the fluorspar district; and log 8 was used by Tippie (1945, p. 1657). Logs 28 and 29 in figure 6 are given by Logan (1931, pp. 461, 466). Section 30 in figure 6 is from Malott (1925, p. 125). Sections 32, 33, and 34 were measured by Swann.[3] All other logs used in the cross sections are sample studies by Atherton, electric logs interpreted by Swann, or composite sample and electric logs.

### STRATIGRAPHY

Inasmuch as there is general agreement in the correlation and nomenclature used within the Eastern Interior Basin for units above the Golconda formation, the uppermost formation which is considered here is the Hardinsburg sandstone which overlies the Golconda. Stratigraphic units are described in descending order. No attempt is made to place the units in formations, groups, or series, and the text headings are not intended to imply stratigraphic rank.

#### HARDINSBURG SANDSTONE

The Hardinsburg sandstone has its maximum development in an area extending from Hamilton and Lawrence counties, Illinois, on the northwest to the outcrop belt through Crittenden and Christian counties, Kentucky, on the south, and to Daviess County, Kentucky, on the east. In this area it lies unconformably on the Golconda formation. Local relief of the pre-Hardinsburg surface is as much as 60 feet in a few localities; channels 10–30 feet deep are quite common (figs. 4, 5, and 6). These narrow channels may be traced for only short distances and appear to die out at either end and do not seem to be connected into valley systems. The thickest known section of Hardinsburg (more than 200 feet) includes about 50 feet of sandstone filling such a channel. Within its region of greatest development the formation is typically 50–70 feet thick; but at a few localities in this same area only 20 feet of gray and varicolored shale with silty laminae separate the massive Glen Dean and Golconda limestones.

Most of the Hardinsburg formation consists of very fine and fine angular light-gray sandstone, greenish siltstone, and gray to black shale. These rocks are present in varying proportions and different successions in localities within a few hundred feet of each other. In the region of its maximum thickness the sandstone is separated from the overlying Glen Dean limestone at almost all localities by a few feet of greenish to dark-gray shale. This shale above the Hardinsburg and below the massive limestone beds of the Glen Dean becomes varicolored and contains dolomite and limestone lenticles as it is traced away from the region of thick Hardinsburg. Within the general region of thick Hardinsburg there are many disconnected areas in which soft gray and greenish shales as much as 30 feet thick, containing occasional red streaks and brown sublithographic dolomite lenses, lie between undisputed Hardinsburg sand-

---

pany, Fluorspar Division, kindly released the information on which composite log 7 in fig. 4 is based. J. H. Steinmesh, president of Minerva Oil Company, kindly released logs 1, 3, 9, and 10 of fig. 4.

[3] Field notes, 1947.

stone and massive Golconda limestone (figs. 4, 5, and 6). To the west, north, and northeast this shale becomes more persistent and includes more dolomite and more red, green, and purplish shales. In these peripheral areas this lower shale is overlain in apparent conformity by the feather edge of the typical Hardinsburg sandstone-siltstone–dark-shale unit. In the parts of the Illinois Basin farthest from the Hardinsburg center, the sandy facies is not developed and the massive Glen Dean and Golconda limestones are separated by only 12–40 feet of varicolored shales containing dolomite and occasional limestone lenses. The sandstone-siltstone–dark-shale unit extends to the outcrop along the southern and eastern margins of the basin, but, in general, only varicolored shale is present at the western, northern, and northeastern borders.

The general practice among oil companies is to place the boundaries of the Hardinsburg formation at the base of the massive Glen Dean limestone and at the top of massive thick-bedded, light-colored oölitic or crystalline limestone of Golconda age, thus including the varicolored shales and brownish dolomites as well as the sandy phase in the Hardinsburg formation (Dana and Scobey, 1941, p. 876). The formation boundaries as thus designated are probably contemporaneous over most or all of the basin and follow horizons which can be identified consistently both in the area of maximum development where channel cutouts occur and in the peripheral areas of apparent conformity. If this practice is followed, the episode or episodes of channel-cutting are placed in the middle rather than at the beginning of Hardinsburg time, since it is apparent that the unconformable surface lies above both the shale-and-dolomite facies and the massive Golconda limestone.

If the name "Hardinsburg" is restricted to the sandy facies (and this has historical precedence), the time of channel-cutting is placed at or near the beginning of Hardinsburg time, and the formation name is limited to genetically related lithologies. Varicolored shale above the sandy phase of the Hardinsburg was originally placed in the Glen Dean formation (Butts, 1917, p. 97). Because it weathers readily, the shale beneath the Hardinsburg sandstone rarely crops out; sandy strata have been reported to rest directly on massive Golconda limestone wherever the contact is visible. The Illinois Geological Survey places this lower shale-and-dolomite zone in the Golconda formation (Workman, 1940, p. 816; Folk and Swann, 1946, p. 11), though it is recognized that the boundaries of the sandstone facies to which the name "Hardinsburg" is thus restricted may not be contemporaneous throughout the basin.

The Hardinsburg has been described in detail because it presents a number of features common to the lower sandstone units in the Chester and the Ste. Genevieve strata. In the lower beds these features are in part obscured and there are disagreements concerning correlation, whereas in the Hardinsburg there is no correlation problem. In those areas in which the Chester sandstones are coarsest and thickest, they rest on conspicuous surfaces of unconformity which are clean and not deeply weathered and in which valley systems have not been noted. Surrounding these areas are belts in which the sandstones are thinner and finer and lie with apparent conformity between shale beds, many of which are in part red, green, and purple. Still farther from the areas of maximum thickness, the sandstones are replaced entirely by shales, which may be fossiliferous and contain limestone or dolomite beds.

### GOLCONDA FORMATION, UPPER SHALE AND LIMESTONE

At the top of the Golconda formation is the unnamed, gray, green, and red shale interbedded with thin lithographic dolomite, which has been described in the preceding section.

Underlying the unnamed shale unit, or in direct contact with the Hardinsburg sandstone, where the shale has been removed by pre-Hardinsburg erosion, is a prominent limestone, generally called "the Golconda lime" and more specifically described as "the massive upper Golconda limestone" by petroleum geologists. It is equivalent to the entire Golconda formation as recognized in the outcrops of Indiana and west-central Kentucky but to only a part of the formation as recognized in the type area in southern Illinois and in western Kentucky. White, light-gray, and light-tan crystalline or crinoidal limestone and dolomitic limestone, gray to tan oölitic limestone, and gray shaly limestone predominate. Oölitic beds are common in the western part and cherty limestones in the eastern part of the basin. In some localities this member is nearly all limestone, but within a short distance it may grade to 50 per cent or more gray shale. It is 40–50 feet thick in the south-central part of the basin, and it thins slightly but retains its essential limestone character to the eastern and western outcrops. It thins markedly toward the north and becomes very shaly, so that, at its northernmost subsurface occurrences, it can hardly be distinguished from the underlying Indian Springs shale.

### INDIAN SPRINGS SHALE

Underlying the massive upper Golconda limestone in the central parts of the Eastern Interior Basin is a third member of the Golconda, a shale 60–100 feet thick, containing minor amounts of limestone, siltstone, and sandstone. The unit is commonly known only as "the Golconda shale," but a representative of the upper part has been named the "Indian Springs shale." The most common rock in this unit is a weak gray shale which weathers readily to a light-colored or olive mud in outcrops and which caves badly in well borings. Red shale is common toward the top of this zone. Siltstones occur at several levels in southernmost Illinois but farther north are confined to the middle and lower portions. They do not occur in western Illinois. Several types of limestone occur in thin beds or lenses that increase in number and importance from eastern to western Illinois. In the west a 5–15-foot bed of red and yellow mottled, fossiliferous, crystalline limestone is a persistent marker near the middle of the member. The upper portion of this shale becomes so limy at the western edge of the Illinois Basin that its contact with the overlying limestone member is obscure.

In Indiana and Kentucky there is very little limestone in this shale unit, and the lower part is replaced by the sandstone known locally as "Jackson" in the subsurface and "Cypress" on the outcrop, described below as the "Big Clifty sandstone." The upper portion of the shale unit extends over the Big Clifty to the Indiana outcrop belt, where it has been named the "Indian Springs shale" by Malott and Thompson (1920, p. 521).

### BIG CLIFTY SANDSTONE

Sandstone lenses occur sporadically in the lower part of the Golconda (Indian Springs) shale as far west as Marion County, Illinois, 60 miles from the Wabash River. The lenses are rather uncommon in Illinois, but eastward they are more abundant and coalesce into a continuous sandstone unit which apparently has its maximum thickness in a

belt extending through eastern Gibson County, Indiana, and Daviess County, Kentucky. There are places in this area at which the sandstone is more than 100 feet thick and lies on an unconformable surface which cuts into and even through the Beech Creek limestone. This is the most continuous and most prominent sandstone in the lower and middle Chester of the Indiana and central Kentucky outcrop belt. Its maximum outcrop thickness, 55–70 feet, is in Hardin, Breckinridge, and Grayson counties, Kentucky, where it caps the Dripping Springs escarpment (Stouder, 1941, p. 48). The name "Big Clifty" was applied to this sandstone in this area by Norwood (1876, p. 369). The name has been little used because this prominent sandstone has generally been considered equivalent to the Cypress sandstone, which had been previously named in southern Illinois. Northeast of a line extending approximately from Shawneetown, Illinois, through Dixon and Greenville, Kentucky, the Big Clifty is separated from the true Cypress by a sheet of Beech Creek ("Barlow") limestone, continuous except for very narrow channels from which it was removed by pre–Big Clifty erosion. Southwest of this line along the outcrop belt through Christian and Caldwell counties, Kentucky, the Big Clifty may lie directly on the true Cypress. It seems likely that even in this area there may be scattered remnants of the Beech Creek separating the two sandstones, as was suggested by Ulrich (1917, p. 94).

#### BEECH CREEK ("BARLOW") LIMESTONE

A thin, but widespread and readily recognized, limestone is the basal unit of the Golconda formation and has been named the "Barlow" line in the subsurface section of Kentucky. It is a dark-brown to dark-gray limestone which in many localities is mottled. Scattered sand grains and very dark-colored limestone granules are common; in some localities there are true oölites. The limestone is commonly very fossiliferous. It contains a diverse fauna in which small brachiopods and mollusks predominate. Polished sections of cores show numerous small gastropods, but the compact matrix makes the fauna difficult to identify. Other types of limestone may be a associated with the dark-colored dense bed, in most cases overlying it.

The "Barlow" line extends across the basin from east to west, appearing in the lower part of the Okaw limestone in southwestern Illinois and as the Beech Creek limestone of Indiana. Unlike other Chester limestones, it thickens consistently toward the north. In the lower Wabash Valley its thickness averages about 8 feet, but it is 20–30 feet thick in the northernmost parts of the basin, where it is by far the most prominent Chester limestone.

The "Barlow" thins southward until it cannot be recognized in many wells in southern Webster and Hopkins counties, Kentucky, or in Saline, Williamson, and even Franklin counties, Illinois. It is present in possibly half the diamond-drill cores from the Illinois-Kentucky fluorspar district as described by trained geologists familiar with its appearance. It is seldom noted in other types of subsurface records from this area and has not been recognized in the poorly exposed outcrops of the lower Golconda shales. In the diamond-drill core used for log 1 of figure 4 it is represented by less than 3 feet of dark, very shaly limestone.

The light-colored, coarsely crystalline limestone containing large crinoid stems which forms approximately the upper three-fourths of the Beech Creek lime-

stone of Indiana can be traced for some distance as the upper portion of the "Barlow," but it is much less persistent than the dark-brown basal portion.

The base of the "Barlow" is the most widely used horizon for contour maps of Illinois Basin areas. It is now generally considered the base of the Golconda formation.

#### CYPRESS SANDSTONE

The Cypress sandstone has its greatest thickness in the south-central part of the Illinois Basin, where it may be more than 200 feet thick. Because in this area it rests on other similar sandstones, its true thickness may never be known. It is predominantly white fine- to medium-grained sandstone and gray siltstone and shale. One or two thin coal beds occur sporadically near the top of the formation. Lateral changes from sandstone to shale are very common. The top part, 30–50 feet thick, is commonly more shaly than the lower part of the formation, but beds of sand may occur at any position up to the very top. A zone of red and green shale and green siltstone occurs at the top of the Cypress throughout a wide belt surrounding the region of greatest thickness. Along the extreme eastern and western borders of the basin the entire unit may be replaced with varicolored shale. Although the uppermost shale zone has been placed in the Golconda formation (Brokaw, 1916, pl. 1; Workman, 1940, p. 216), the current tendency is to include it with the Cypress, as it is commonly interbedded with thin sandstone beds. The varicolored shales equivalent to the true Cypress have been named "Ruma" (S. Weller, 1913, p. 126) in southwestern Illinois and "Elwren" (Malott, 1919, p. 11) in Indiana. The Cypress rests unconformably on lower formations, except possibly in the most northern and eastern parts of the basin.

#### PAINT CREEK (RIDENHOWER) FORMATION

Approximately midway between the persistent Beech Creek ("Barlow") limestone and the equally persistent Downeys Bluff limestone, there occurs a discontinuous zone which is composed of diverse rock types and has a maximum thickness of 100 feet. This zone has been described as three formations—the Reelsville limestone, Sample sandstone, and Beaver Bend limestone in Indiana (Malott, 1919, pp. 9–11); as a single formation—the Ridenhower shale in southeastern Illinois (Butts, 1917, p. 73); and as part of the Paint Creek formation in southwestern Illinois. Although single units within this zone may be traced a number of miles, it is difficult to characterize the zone as a whole. Several factors have shared in producing the complex stratigraphy of this zone. Pre-Cypress erosion removed the zone entirely in some localities and reduced its thickness in many areas. The Cypress sediments deposited on the eroded surface are, in some places, similar to the beds removed. There is at least one pronounced unconformity within the zone. Rapid lateral and vertical changes in deposition can be demonstrated in a number of single outcrops.

There is a belt 10–20 miles wide extending north-northeast from the vicinity of Marion and Uniontown, Kentucky, toward Washington, Indiana, in which the entire column between the "Barlow" and the Downeys Bluff is occupied by sandstone (fig. 5). It is probable that the Indiana classification of three units can be followed rather consistently between the east side of this belt and the region in south-central Kentucky where all clastics disappear from the section. Available data from the belt of thick sandstone do not show any clear break between the

pre-Cypress unconformity and the base of the Bethel, so it is not certain that both Bethel and Sample sandstones are present, although this seems probable. It also is not evident whether the Beaver Bend and Reelsville limestones were deposited and later eroded from this area or were never deposited here.

In the subsurface section of Illinois west of the continuous sandstone belt, the application of a threefold division of the Paint Creek above the Bethel is much more difficult than in Kentucky and Indiana. The division is suggested in some logs shown in figure 3. Certain geologists who have studied intensively the area north of logs 30–50 of this section recognize beneath the pre-Cypress unconformity a "gray limestone" (Reelsville?) which has generally been removed by erosion, a shale and micaceous sandstone zone (Sample?), a "yellow limestone" (Beaver Bend?), a blastoidal shale, and the Bethel. The unconformity cuts across all these strata, so that the Cypress may rest directly on the Bethel. The writers are unable to distinguish these zones in their work in Lawrence and Edwards counties, Illinois; therefore, they recognize only a single zone, which consists primarily of dark greenish-gray shale with variable proportions of fine-grained calcareous sandstone and sandy fossiliferous, partly oölitic limestone. In western Illinois this zone becomes more calcareous and is represented in the Paint Creek outcrop belt by as much as 50 feet of limestone.

### BETHEL SANDSTONE

The Bethel sandstone has its greatest thickness east of the area of maximum Cypress sandstone, locally measuring 100 feet or more. It contains some quartz-pebble conglomerate, but the formation usually consists of more or less calcareous light-gray, very fine to fine-grained sandstone with some medium-grained sandstone. Dark-gray to green shale containing laminae of siltstone is quite common. The type locality near Marion, Crittenden County, Kentucky, is probably within the belt of continuous Bethel-Cypress or Bethel-Sample-Cypress sandstone described above (Butts, 1917, p. 63; 1929, p. 46).

To the north and west of the area of maximum thickness the formation becomes thinner, more shaly, more calcareous, and fossiliferous. The writers know of only a single outcrop of the Bethel in southwestern Illinois. It is a calcareous fossiliferous sandstone a few feet above the deep-red nonlaminated shale member of the Paint Creek formation in the NE. corner of the SE.$\frac{1}{4}$ and NE.$\frac{1}{4}$ of Sec. 23, T. 5 S., R. 9 W., about 2 miles east of Prairie du Rocher, Randolph County. This locality was noted by Stuart Weller in his report on the "Geology of Parts of Monroe and Randolph Counties."[4] In the rest of the southwestern Illinois outcrop area the position of the Bethel is occupied by varicolored calcareous shales that in some localities are highly fossiliferous.

The sandstone phase of the Bethel continues to the eastern outcrop, where it is known as the "Mooretown sandstone" (Cumings, 1922, p. 515) and includes thin but persistent coal streaks overlain by a few feet of dark-gray shale. The "limestone" indicated in the Bethel position in log 22 of figure 6 is probably calcareous sandstone; several sandstones in this well had abnormally high apparent resistivity, and samples were not available for comparison. As is true with most Chester sandstone units, the Bethel is recognized by its relation to the as-

[4] Illinois Geological Survey, unpublished manuscript.

sociated limestone beds rather than by any inherent characteristics.

### DOWNEYS BLUFF LIMESTONE

A persistent limestone which is useful in correlating lower Chester beds has been named the "Downeys Bluff" by F. E. Tippie (Atherton, 1948, p. 129). The type locality is in the Ohio River bluff at Rosiclare, where the limestone underlies the Bethel sandstone and overlies the Shetlerville limestone and shale. In southwestern Illinois these beds consist of bluish calcareous shales with platy, fine-grained limestone layers lying between the deep-red nonlaminated shale member of the Paint Creek and the Yankeetown chert. This zone was described by S. Weller (1913, p. 125; 1920*a*, p. 294) as the basal member of the Paint Creek formation. Two other kinds of rock are found in this zone in the western outcrop belt. One is a yellowish, crystalline, crinoidal limestone. The other is a characteristic light-gray, coarsely crystalline, fossiliferous, and somewhat sandy limestone in which numerous blastoid and crinoid plates are in part replaced by a bright-pink or salmon-colored chert. Replacement of colored crinoid plates by pink or red chert is not known in other lower Chester formations. The gray limestone with pink chert fossils can be seen in the outcrops along a branch of Carr Creek in the NE.$\frac{1}{4}$, Sec. 33, T. 1 S., R. 10 W., about 2 miles south of Columbia, Monroe County, Illinois. It is recognized throughout the subsurface section of southwestern and south-central Illinois, and the entire Downeys Bluff is known by the informal name "Pink Crinoidal" throughout Illinois. In eastern Illinois the bulk of the Downeys Bluff is a light-brown to gray crinoidal limestone, the upper part cherty and the lower part slightly sandy. Pink to red chert is reported in some wells in eastern Illinois and even in Indiana (Dana and Scobey, 1941, p. 879). Throughout most of eastern Illinois and western Kentucky and Indiana the Downeys Bluff may be recognized by a very characteristic double-peaked resistivity curve on the electric log. The saddle has been interpreted in figures 5 and 6 as a thin shale, although it may be an argillaceous limestone. The Downeys Bluff is represented in the Indiana outcrop by the upper part of the Paoli limestone.

### "BENOIST" (YANKEETOWN) SANDSTONE

All the sandstones considered so far are prominent beds in the eastern or south-central parts of the Eastern Interior Basin; on the southwestern Illinois outcrop they are represented by, at most, a few feet of impure sandstone. In contrast, Chester and Ste. Genevieve sandstones below the Downeys Bluff limestone (with a possible minor exception) are best displayed near the western margin of the basin.

The uppermost of these western sandstones has been named the "Benoist" sandstone in Sandoval oil field, Marion County, Illinois, about 15 miles north of Boyd. The stratigraphic section at Sandoval is very similar to that shown in figure 3, log 1. In this area the "Benoist" is separated from the "Pink Crinoidal" or Downeys Bluff limestone by a very few feet of green shale. Continuous cores show that immediately above or interbedded in the top foot or two of the "Benoist" are thin beds (nodules?) of a brown siliceous limestone, which grades into, or is replaced by, chert. Westward this zone at the top of the "Benoist" can be traced into at least some of the outcrops of the Yankeetown chert (Tippie, 1943, p. 141). Swann agrees with Weller and Sutton (1940, p. 826, ftn. 19) that

the Yankeetown includes residual chert from several horizons, and therefore he hesitates to apply the outcrop term to the sandstone. The "Benoist" is typically 40–60 feet thick where best developed, but it rests directly on continuous Renault or Aux Vases sandstone in many localities where its true thickness cannot be determined.

Eastward from its area of maximum thickness the "Benoist" thins rapidly, and in easternmost Illinois it consists only of about 20 feet of dark-gray, green, and red shale, in the upper part of which occur lenses of calcareous light-gray, red, or greenish fine-grained sandstone or siltstone. The lower part of this shale is interbedded with layers of brownish-gray argillaceous, fossiliferous limestone and grades downward into the main body of Shetlerville limestone. The entire sequence was named the "Shetlerville formation" by S. Weller (1920*a*, p. 290), and the shale and the sandstone are now called "Middle Renault" by most operators in the Wabash Valley oil fields. East of the Wabash the sandstone is absent. Its position is indicated by 2–3 feet of fossiliferous shale near the middle of the Paoli limestone at the more northern of the Indiana outcrops; but even this shale is lacking in the southernmost Indiana and Kentucky occurrences of the Paoli.

#### SHETLERVILLE LIMESTONE

The lower portion of the Shetlerville formation is predominantly limestone, grayish-brown in color, oölitic, fossiliferous, argillaceous, and in many places sandy. Oölites with dark centers are common; and red, pink, olive, and yellow oölites occur. The Shetlerville in Illinois is quite impure, as is indicated by the insoluble residues described by Tippie (1944, p. 157); but it apparently becomes purer in Indiana and Kentucky, where it forms the lower portion of the Paoli limestone. It makes up the bulk of the "Lower Renault" limestone of the oil fields in the deeper part of the basin.

#### LEVIAS LIMESTONE

The Levias limestone of the fluorspar district is a high-purity oölitic or crinoidal limestone containing coarse pink crinoid fragments and, at its base, minor amounts of sand. It may be traced northward from the outcrop to Lawrence County, Illinois, as a thin, light-gray or pink oölite at the base of the "Lower Renault" (Shetlerville) limestone. It is not known to be more than 15 feet thick in the subsurface of the basin and is absent from many localities. As this bed contains columnals of *Platycrinus penicillus* in the outcrop area, it is placed in the Ste. Genevieve formation. However, it overlies a continuous sandstone horizon which can be traced north and west to a point where nearly 200 feet of sandstone and shale are present beneath what appears to be the position of the type Levias (figs. 3 and 4).

#### ROSICLARE ("BASIN AUX VASES") SANDSTONE

The Rosiclare sandstone of the fluorspar outcrop district of southeastern Illinois and western Kentucky is an extremely fine-grained sandstone which approaches siltstone in grain size. It continues to the north with similar lithology and is productive in many oil fields in Hamilton, Wayne, and White counties in southeastern Illinois, where it is called the "Aux Vases" sandstone. It is sometimes distinguished from the coarser-grained sandstones of western Illinois by the title "basin Aux Vases." Because of its fine grain size and resultant high connate water retention, it is characterized by abnormally low electrical resistivity

## TABLE 1

**WELLS USED IN CROSS SECTION BETWEEN BOYD OIL FIELD, JEFFERSON COUNTY ILLINOIS, AND THE VICINITY OF NEW HAVEN IN THE WABASH VALLEY (FIG. 3)**

| Index No. | State, County (Illinois) | Operator | No. and Farm | Spot or Footage | Sec., T., R. |
|---|---|---|---|---|---|
| 1 | Jefferson | D. Schwab *et al.* | 1 Hutchings-Haldorson | NE. SE. NW. | 36-1S.-1E. |
| 2 | Jefferson | T. B. Dirickson | 1 Miller | NE. NE. NE. | 18-2S.-2E. |
| 3 | Jefferson | E. J. Ruwaldt | 1 W. J. Hynes | NW. SW. NW. | 28-2S.-2E. |
| 4 | Jefferson | A. W. Gerson | 1 W. B. Horton | SE. SE. NW. | 27-2S.-2E. |
| 5 | Jefferson | Magnolia Pet. Co. | 1 Bullock Unit | NW. NW. SE. | 34-2S.-2E. |
| 6 | Jefferson | H. H. Wegener | 1 Grant Comm. | W. NW. SE. | 35-2S.-2E. |
| 7 | Jefferson | N. Redwine | 2 Howard-Casey | SE. NE. | 6-3S.-3E. |
| 8 | Jefferson | N. Redwine | 1 K. Gee | NE. SW. SE. | 6-3S.-3E. |
| 9 | Jefferson | Magnolia Pet. Co. | 1 Badgett | NE. NW. SE. | 7-3S.-3E. |
| 10 | Jefferson | Magnolia Pet. Co. | 1 Daniels Unit | SW. SW. SW. | 8-3S.-3E. |
| 11 | Jefferson | W. I. Lewis | 1 Schul | SE. SE. NW. | 9-3S.-3E. |
| 12 | Jefferson | Texas Co. | 1 N. Cowger | NW. NW. NW. | 11-3S.-3E. |
| 13 | Jefferson | J. V. Canterbury | 1 L. S. Kent | NW. NE. SE. | 12-3S.-3E. |
| 14 | Jefferson | Canterbury & Gill | 1 R. Ross | SE. SE. SE. | 7-3S.-4E. |
| 15 | Jefferson | Phillips-Gussman | 1 R. E. Sheppard | C. SE. SE. | 8-3S.-4E. |
| 16 | Jefferson | B. Martin *et al.* | 1 Rittermeyer | SE. SW. NE. | 9-3S.-4E. |
| 17 | Jefferson | C. E. Brehm | 1 Burnett-Shelton Comm. | SE. SW. SW. | 3-3S.-4E. |
| 18 | Jefferson | Gulf Refining Co. | 1 Homer | SW. SE. NW. | 3-3S.-4E. |
| 19 | Jefferson | Kewanee Oil & Gas | 2 Derringer | N. NE. SE. | 2-3S.-4E. |
| 20 | Jefferson | Tidewater Assoc. Oil Co. | "B"-1 Newton Investment Co. | N. NE. SW. | 1-3S.-4E. |
| 21 | Jefferson | Magnolia Pet. Co. | 1 F. R. Johnson | N. NE. SE. | 1-3S.-4E. |
| 22 | Wayne | Lario Oil & Gas | 1 R. Dodge | E. NE. NE. | 7-3S.-5E. |
| 23 | Wayne | C. Crosby & Witt | 1 Murphy | E. SW. NE. | 15-3S.-5E. |
| 24 | Hamilton | Seaboard | 1 G. Knapp | N. SW. SW. | 22-3S.-5E. |
| 25 | Hamilton | Seaboard | 2 D. Garrison | N. SW. NW. | 27-3S.-5E. |
| 26 | Hamilton | Gulf Refining Co. | 1 F. Zellers | SE. NW. SE. | 27-3S.-5E. |
| 27 | Hamilton | Exchange Oil Co. | "A"-1 E. Silliman | SW. NW. NW. | 35-3S.-5E. |
| 28 | Hamilton | Oil Carriers | 1 T. Rose | SW. NE. SE. | 36-3S.-5E. |
| 29 | Hamilton | Magnolia Pet. Co. | "A"-1 Kaufman | SE. SE. NW. | 7-4S.-6E. |
| 30 | Hamilton | Magnolia Pet. Co. | 1 Haas (Heil) | NW. NW. SW. | 16-4S.-6E. |
| 31 | Hamilton | Oil Management *et al.* | 1 Leach & Gilpin | C. SW. NE. | 14-4S.-6E. |
| 32 | Hamilton | Phillips Pet. Co. | 1 Holla | NW. NW. NE. | 13-4S.-6E. |
| 33 | Hamilton | Phillips Pet. Co. | 1 Wilma | SE. SW. NW. | 7-4S.-7E. |
| 34 | Hamilton | Texas Co. | 3 Flannigan | NE. SE. SW. | 17-4S.-7E. |
| 35 | Hamilton | Texas Co. | 2 Minton Comm. | SW. NE. NE. | 20-4S.-7E. |
| 36 | Hamilton | Texas Co. | 3 S. Minton | NE. NW. SW. | 21-4S.-7E. |
| 37 | Hamilton | Oldfield & Spires | 2 York | NE. SE. SE. | 22-4S.-7E. |
| 38 | Hamilton | Oil Management *et al.* | 1 J. Keith | NW. NW. SE. | 26-4S.-7E. |
| 39 | Hamilton | Kingwood Oil Co. | 1 Swadier | NE. NW. SE. | 2-5S.-7E. |
| 40 | Hamilton | Kingwood Oil Co. | 1 McGuire | W. SW. NW. | 11-5S.-7E. |
| 41 | Hamilton | Lomelino & Williamson | 1 Biggerstaff | NE. SE. NW. | 14-5S.-7E. |
| 42 | Hamilton | T. Harvey | 1 Wilson | NE. SE. SW. | 23-5S.-7E. |
| 43 | Hamilton | B. M. Heath | 1 Klemm Heirs | C. NE. SE. | 35-5S.-7E. |
| 44 | White | National Assoc. Pet. Co. | 1 A. Dalby | SE. SW. NE. | 6-6S.-8E. |
| 45 | White | Gossett & Swensen | 1 Justice | SE. NW. NE. | 20-6S.-8E. |
| 46 | White | Luttrell | 1 T. Aud | NE. NE. SE. | 21-6S.-8E. |
| 47 | White | J. Reznik | 1 Mills | NE. SW. NW. | 35-6S.-8E. |
| 48 | White | Sinclair-Wyoming | 1 A. Cobbell | SW. SW. SW. | 36-6S.-8E. |
| 49 | White | Gulf Refining Co. | 1 J. Moore | SW. SW. NW. | 12-7S.-8E. |
| 50 | White | Kingwood Oil Co. | 1 P. Martin | NW. SW. NE. | 13-7S.-8E. |
| 51 | White | Sinclair-Wyoming Oil Co. | 1 C. E. Wilson | SW. NE. SE. | 18-7S.-9E. |
| 52 | Gallatin | Skelly *et al.* | 1 H. H. Hale | NE. NE. NE. | 21-7S.-9E. |
| 53 | Gallatin | Murphy Oil *et al.* | 1 Spence | N. NE. NW. | 33-7S.-9E. |
| 54 | Gallatin | N. V. Duncan | 1 Greer | SE. SE. SE. | 35-7S.-9E. |
| 55 | Gallatin | R. L. Kinkaid | 1 A. B. Schmidt | NE. NE. NW. | 33-7S.-10E. |
| 56 | Gallatin | Hageman & Pond | 1 Stofleth-Cokes | NE. NE. NW. | 34-7S.-10E. |

for Illinois Basin sandstones. This electrical characteristic is indicated both on logs run in the basin and on a log run on a typical fluorspar exploration hole in Hardin County, Illinois. The sandstone is calcareous and variable in color; red sandstone is common where the formation is not productive, but the productive sand is very light colored, nearly white.

The Rosiclare shows considerable lateral changes within a very short distance and includes numerous lithologic types —siltstone, shale, sandy limestone, and dolomite—in addition to sandstone. Some sandstone beds contain very small, light-brown, round limestone grains and grade laterally to sandy oölitic limestone. Hematite-ringed oölites are characteristic of the zone, and a basal detrital conglomerate is present at many points. The Rosiclare ("basin Aux Vases") sandstone of southeastern Illinois appears, at least superficially, to grade imperceptibly into the thicker and coarser Aux Vases sandstone of the oil fields on the western flank of the basin, as is indicated by figure 3. The possibility of a transgressive overlap of western coarse-grained Aux Vases sandstone on eastern fine-grained "basin Aux Vases" is recognized; but the evidence supporting this interpretation is not conclusive, and further study is necessary.

TABLE 2

WELLS AND OUTCROPS USED IN CROSS SECTION FROM GOLCONDA, ILLINOIS TO NEW HAVEN, ILLINOIS (FIG. 4)

| Index No. | State and County | Operator | No. and Farm | Spot or Footage | Sec., T., R. |
|---|---|---|---|---|---|
| 1 | Ky., Livingston | Minerva Oil Co. | DD 1 Robinson | 7,200 WL., 10,550 NL. quad. | 7-J-14 |
| 2 | Ill., Pope | Outcrop at Golconda | | SE. NW. | 30-13S.-7E. |
| 3 | Ky., Livingston | Minerva Oil Co. | 2 O. Morton | 5,150 WL., 7,550 SL. quad. | 17-K-14 |
| 4 | Ill., Pope | Outcrop at Rock Quarry School | | SE. | 5-13S.-7E. |
| 5 | Ill., Hardin | Outcrop at Shetlerville | | NE. SW. | 35-12S.-7E. |
| 6 | Ill., Hardin | Outcrop at Downeys Bluff (Rosiclare) | | NW. SE. | 5-13S.-8E. |
| 7 | Ill., Hardin | Rosiclare Lead & Fluorspar Mining Co. | Ac 2 Fee | SE. SE. NW. | 32-12S.-8E. |
| | | Hillside Fluorspar Mines | 107-D Fee | NW. NW. SE. | 29-12S.-8E. |
| 8 | Ill., Hardin | Victory Fluorspar | 126 Fee | NE. SE. | 33-11S.-9E. |
| 9 | Ill., Hardin | Minerva Oil Co. | 1 C. M. Austin | SW. SW. NE. | 26-11S.-9E. |
| 10 | Ill., Hardin | Minerva Oil Co. | 7 U.S. Forest Service | SE. NE. SE. | 24-11S.-9E. |
| | | Minerva Oil Co. | 7 Milligan | CW. line, NW. NE. SW. | 19-11S.-9E. |
| 11 | Ky., Union | H. H. Weinert | 1 Union Trust Co. | N.½ N.½ N.½ | 22-O-18 |
| 12 | Ky., Union | Sohio Prod. Co. *et al.* | 2 Boswell | SE. | 14-O-18 |
| 13 | Ill., Gallatin | Cherry & Kidd | 1 Gray | NE. NE. SE. | 15-9S.-10E. |
| 14 | Ill., Gallatin | Sinclair-Wyoming | 1 E. Hines | SE. SE. NW. | 3-9S.-10E. |
| 15 | Ill., Gallatin | Sohio Prod. Co. | 2 Nat. Resources | SW. SW. SW. | 28-8S.-10E. |
| 16 | Ill., Gallatin | Carter | 2 Jordan | NE. SE. NE. | 21-8S.-10E. |
| 17 | Ill., Gallatin | Cherry & Kidd | 12 Kerwin | SW. SW. SW. | 11-8S.-10E. |
| 18 | Ill., Gallatin | Hagemann & Pond | 1 Stofleth-Cokes | NE. NE. NW. | 34-7S.-10E. |

### FREDONIA LIMESTONE

The Fredonia limestone of the fluorspar district includes beds equivalent to

the entire sequence called "Ste. Genevieve" in the eastern Illinois oil fields (fig. 4). The Fredonia is mainly limestone, oölitic, slightly sandy, light-gray, buff, or brown, and medium to coarse textured. Porous oölitic zones are known as "McClosky," although the upper two such zones are designated the "Lower O'Hara pay" and the "Rosiclare pay" by many geologists in the eastern Illinois oil fields. Many beds may be traced accurately over short distances, but long-range correlations within the mass of limestone are very difficult because of rapid lateral variation and the recurrence of similar lithologic types at different levels in the section.

Sandy zones occur at many horizons and may grade laterally into sandstone. One such sandy zone has been named the "Spar Mountain sandstone" by Tippie (1945, p. 1658) at a locality about 1 mile

TABLE 3

WELLS USED IN CROSS SECTION FROM VICINITY OF MAUNIE, ILLINOIS, TO POOLE OIL-FIELD AREA, HENDERSON CO., KY. (FIG. 5)

| Index No. | State and County | Operator | No. and Farm | Spot or Footage | Sec., T., R., |
|---|---|---|---|---|---|
| 1... | Ind., Posey | Gulf Refining Co. | 1 Aldrich Community | SE. NE. NE. | 8-6S.-14W. |
| 2... | Ind., Posey | Paul Maier | 1 Aldrich | C. NE. NE. | 4-6S.-14W. |
| 3... | Ind., Posey | Morgenstern Oil Co., Inc. | 1 Mentzer | NW. SW. NE. | 34-5S.-14W. |
| 4... | Ind., Posey | Justrite Drilling Co. | 1 French | CW.½ W.½ W.½ | 1-6S.-14W. |
| 5... | Ind., Posey | B. M. Heath | 1 Noble Utley | SE. NE. SW. | 30-5S.-13W. |
| 6... | Ind., Posey | Milton A. Lobree<br>C. E. O'Neal *et al.* | 1 W. Jackson | NW. NW. NW. | 27-5S.-13W. |
| 7... | Ind., Posey | Paul Rossi | 1 Kincheloe & Williams | NE. NE. NW. | 35-5S.-13W. |
| 8... | Ind., Posey | Gulf Refining Co. | 3 Lang | SE. NE. SE. | 6-6S.-12W. |
| 9... | Ind., Posey | Nelson Development Co. | 1 Horstman | NE. SE. NE. | 7-6S.-12W. |
| 10... | Ind., Posey | Gulf Refining Co. | 1 Reineke | NE. SE. NW. | 8-6S.-12W. |
| 11... | Ind., Posey | L. E. Butzman *et al.* | 1 B. R. Junker | NW. NW. NW. | 9-6S.-12W. |
| 12... | Ind., Posey | S. C. Yingling | 1 Eckhoff | SE. NW. SE. | 21-6S.-12W. |
| 13... | Ind., Posey | Ward-Larson | 1 Eickhoff-Wolf | SW. NE. SW. | 22-6S.-12W. |
| 14... | Ind., Posey | Fleming & A. K. Swann | 1 John Hartmann | NW. NW. SE. | 27-6S.-12W. |
| 15... | Ind., Posey | E. T. Wix | 1 E. Miller | NW. SE. SW. | 26-6S.-12W. |
| 16... | Ind., Vanderburg | Justrite Drilling Co. | 1 Miller | N.½ SE. NW. | 31-6S.-11W. |
| 17... | Ind., Vanderburg | Roy Lee, Trustee | 1 Oakland City College | E.½ SE. SW. | 3-7S.-11W. |
| 18... | Ind., Vanderburg | Sun Oil Co. | 1 Adcock Unit | NE. SE. SE. | 23-7S.-11W. |
| 19... | Ind., Vanderburg | C. E. O'Neal & Co. | 1 Kuester | NE. SW. NE. | 3-8S.-11W. |
| 20... | Ind., Vanderburg | Calvert-Willis & Delta | 1 Simmons | NW. NW. SE. | 10-8S.-11W. |
| 21... | Ky., Henderson | Basin Drilling Co. | 5 Carl Smith | 8,890′ NL., 9,810′ EL. quad. | 8-P-23 |
| 22... | Ky., Henderson | Sohio Producing Co. | 1 Bartley | 6,900′ SL., 7,950′ WL. of quad. | 17-P-23 |
| 23... | Ky., Henderson | Herndon Drilling Co. | 1 Barrett | 2,900′ SL., 11,250′ EL. quad. | 23-P-23 |
| 24... | Ky., Henderson | Carter Oil Co. | 1 Ben Rudy | 4,500′ NL., 11,250′ WL. quad. | 3-O-23 |
| 25... | Ky., Henderson | W. F. Bilsky | 1 F. P. Royster | 13,600′ SL., 12,300′ EL. quad. | 13-O-23 |
| 26... | Ky., Henderson | Sohio Producing Co. & W. E. Hupp | 1 Minton | 6,800′ SL., 7,500′ WL. quad. | 17-O-23 |
| 27... | Ky., Henderson | Carter Oil Co. | 6 S. T. Denton | 1,290′ SL., 5,670′ EL. quad. | 22-O-23 |
| 28... | Ky., Henderson | Sohio Producing Co. | 1 O. Royster 95A. | 3,150′ SL., 400′ WL. quad. | 25-O-24 |

south of the well illustrated in log 8 in figure 4. The sequence of minor lithologic types above the sandstone at this point is very similar to that above the "Rosiclare" in certain of the southeasternmost Illinois oil fields. In other areas different sand zones appear more prominent, and it is probable that different zones are called "Rosiclare" in different parts of the basin. Sandstones are much more common to the west, and in certain areas as many as three thick "Rosiclare" sandstones can be recognized, reaching 100 feet in total thickness. One moderately thick sandstone is indicated at the western edge of figure 3.

TABLE 4

WELLS AND OUTCROPS USED IN CROSS SECTION FROM POOLE OIL-FIELD AREA HENDERSON CO., KY., TO LEAVENWORTH, CRAWFORD CO. INDIANA (FIG. 6)

| Index No. | State and County | Operator | No. and Farm | Spot or Footage | Sec., T., R. |
|---|---|---|---|---|---|
| 1.... | Ky., Henderson | Sohio Producing Co. | 1 O. Royster 95A. | 3,150 SL., 400 WL. quad. | 25-O-24 |
| 2.... | Ky., Henderson | Cherry & Kidd | 1-A L. Eakins | 300 EL., 200 NL. sec. | 4-N-24 |
| 3.... | Ky., Henderson | Cherry & Kidd | 1 T. J. Pritchett | 1,050 EL., 2,950 SL. sec. | 19-Q-2[illegible] |
| 4.... | Ky., Henderson | Sohio Producing Co. | 1 H. J. Knight | 8,400 SL., 550 EL. quad. | 20-[illegible]-24 |
| 5.... | Ky., Henderson | Ryan Oil Co. | 1 V. Crafton | 3,700 SL., 1,600 EL. sec. | 9-O-24 |
| 6.... | Ky., Henderson | National Assoc. Pet. Co. | 1 Williams | 3,300 NL., 4,470 EL. sec. | 1-O-24 |
| 7.... | Ky., Henderson | Carter Oil Co. | 1 H. P. Barrett | 5,880 SL., 4,650 WL. quad. | 25-P-25 |
| 8.... | Ky., Henderson | Kingwood Oil Co. | 1 Jones | 7,380 WL., 6,900 SL. quad. | 17-P-25 |
| 9.... | Ky., Henderson | Reznik | 1 Moss | 1,750 WL., 1,900 SL. sec. | 7-P-25 |
| 10.... | Ky., Henderson | Sinclair Prairie Oil Co. | 1 Hatchett | 12,500 NL., 9,550 EL. quad. | 12-P-25 |
| 11.... | Ky., Henderson | J. H. Williams | 1 J. L. Overby | 8,000 EL., 700 NL. quad. | 2-P-25 |
| 12.... | Ky., Henderson | McCummings Oil Co. | 1 E. Allen | 1,950 SL., 2,300 WL. sec. | 5-P-26 |
| 13.... | Ky., Henderson | Kentucky Producers Corp. | 1 O. Breitschere | 10,250 EL., 100 SL. quad. | 23-Q-26 |
| 14.... | Ky., Henderson | Coaster Co. | 1 Haynes | 5,600 EL., 5,950 SL. quad. | 22-Q-26 |
| 15.... | Ky., Henderson | Farmer & Chenault | 1 R. E. Dunbar | 9,260 SL., 3,240 EL. quad. | 20-Q-26 |
| 16.... | Ky., Henderson | Farmer & Chenault | 1 M. Bruck | 12,950 SL., 1,850 WL. quad. | 15-Q-27 |
| 17.... | Ky., Daviess | Sohio & Hupp Pet. | 3 Reno Heirs | 2,750 NL., 800 EL. sec. | 14-Q-27 |
| 18.... | Ky., Henderson | Farmer & Chenault | 1 Bower & Heppler | 4,400 WL., 9,600 NL. quad. | 6-Q-27 |
| 19.... | Ind., Warrick | W. Chenault | 1 Turner | Approx. C. SW. SE. | 36-6S.-9W. |
| 20.... | Ind., Warrick | Ohio Oil Co. | 1 R. Jones | NE. NE. SW. | 20-6S.-8W. |
| 21.... | Ind., Warrick | Eureka Oil Co. | 1 T. H. Helms | SW. SW. NW. | 23-6S.-8W. |
| 22.... | Ind., Warrick | Sunlight Coal Co. | 1 Hart | SE. SE. NW. | 27-5S.-8W. |
| 23.... | Ind., Warrick | Cherry & Kidd (Ashland Oil) | 1 Verona Coal Co. | SW. SE. SW. | 8-5S.-7W. |
| 24.... | Ind., Warrick | L. T. Phillips | 2 T. S. Phillips | NE. NE. SW. | 32-4S.-6W. |
| 25.... | Ind., Spencer | Texas Co. | 1 Hanning | SW. SW. NE. | 2-5S.-5W. |
| 26.... | Ind., Spencer | Texas Co. | 1 Gogel | SE. NW. SW. | 25-4S.-5W. |
| 27.... | Ind., Spencer | Ohio Oil Co. | 1 Holtzman | NE. SE. SE. | 28-4S.-4W. |
| 28.... | Ind., Perry | Ohio Oil Co. | 1 W. Epple | NE. NE. NE. | 8-5S.-3W. |
| 29.... | Ind., Perry | Ohio Oil Co. | 1 J. C. Harbaville | NE. SE. NW. | 36-4S.-3W. |
| 30.... | Ind., Perry | Outcrop at Branchville | | SW.¼ | 13-4S.-2W. |
| 31.... | Ind., Perry | Sun Oil Co. | 1 Gibson | SE. SE. | 17-4S.-1W. |
| 32.... | Ind., Perry | Outcrop at Courcier Hill | | | 10 & 11, 5S., 1W. |
| 33.... | Ind., Crawford | Outcrop at Sulphur | | SW.¼ | 31-3S.-1E. |
| 34.... | Ind., Crawford | Outcrop at Leavenworth | | E.½ | 6-4S.-2E. |

LOCATIONS USED IN CROSS SECTIONS

Locations in Illinois (see tables 1, 2) and Indiana (tables 2–4) are described by the United States Land Office system of sections, townships, and ranges. Locations in Kentucky (see tables 3, 4) are given in the co-ordinate system for Kentucky originated by the Carter Oil Company about 1937 and now in common use by the oil industry. In this sytem, 5′ quadrangles are lettered in alphabetical order north from latitude 36°30′ and are numbered consecutively east from longitude 89°30′. Each 5′ quadrangle is thus indicated by a letter analogous to the township and a number analogous to the range of the standard system. Each of these quadrangles is subdivided into twenty-five 1′ rectangles which are called "sections" and approximate the mile-square Land Office sections in area but not in shape. The northeastern section in each quadrangle is numbered 1

and the northwestern 5. Six is immediately south of 5, and 10 is south of 1. The southeastern section is 21 and the southwestern is 25.

ACKNOWLEDGMENTS.—A report of this kind is possible only with the co-operation of numerous geologists and officials of oil companies, flourspar companies, and the several state agencies represented in the Eastern Interior region. It is probable that none of the correlations here presented are original; many have been worked out repeatedly and independently by different geologists and are part of the common unpublished knowledge of geologists active in the area. Unpublished reports of F. E. Tippie, formerly with the Illinois Geological Survey, have been used freely. The criticism and suggestions of L. E. Workman, A. H. Bell, H. B. Willman, and other members of the Survey have been very helpful.

## REFERENCES CITED

ATHERTON, ELWOOD (1948) Some Chester outcrop and subsurface sections in southeastern Illinois: Illinois Acad. Sci. Trans., vol. 40, pp. 122–139; Illinois Geol. Survey Circ. 144.

BROKAW, A. D. (1916) Oil investigations in Illinois: parts of Saline, Johnson, Pope, and Williamson counties: preliminary extract, Illinois Geol. Survey Bull. 35, pp. 1–13, 3 pls.

BUTTS, CHARLES (1917) Descriptions and correlations of the Mississippian formations of western Kentucky: Kentucky Geol. Survey, ser. 5, vol. 1, pp. 1–119.

——— (1929) Some issues in Chester stratigraphy in Kentucky and Illinois: Jour. Geology, vol. 37, pp. 30–46.

COOPER, C. L. (1941) Chester ostracodes of Illinois: Illinois Geol. Survey, Rept. Inv. 77, pp. 1–101, 14 pls.

CUMINGS, E. R. (1922) The nomenclature and description of the geological formations of Indiana: Indiana Dept. Cons. Pub. 21, pp. 403–571.

DANA, P. L., and SCOBEY, E. H. (1941) Cross section of Chester of Illinois basin: Am. Assoc. Petroleum Geologists Bull. 25, pp. 871–882.

FOLK, S. H., and SWANN, D. H. (1946) King oil field, Jefferson County, Illinois: Illinois Geol. Survey Rept. Inv. 119, pp. 1–27.

LOGAN, W. N. (1931) The sub-surface strata of Indiana. Indiana Dept. Cons. Pub. 108, pp. 1–790.

MALOTT, C. E. (1919) The "American Bottoms" region of eastern Greene County, Indiana—a type unit in southern Indiana physiography: Indiana Univ. Studies, vol. 6, Study 40, pp. 1–61.

——— (1925) The upper Chester of Indiana: Indiana Acad. Sci. Proc., vol. 34, pp. 103–132.

——— (1931) Geologic structure in the Indian and Trinity Springs locality, Martin County, Ind.: Indiana Acad. Sci. Proc., vol. 40, pp. 217–231.

——— (1946) The geology of Cataract Falls, Owen County, Indiana: Jour. Geology, vol. 54, pp. 322–326.

———, and THOMPSON, J. D., JR. (1920) The stratigraphy of the Chester series of southern Indiana (abstr.): Science, new ser., vol. 51. pp. 521–522.

NORWOOD, C. J. (1876) Report on the geology of the region adjacent to the Louisville, Paducah, and Southwestern Railroad: Kentucky Geol. Survey Rept. Prog. vol. 1, new ser., pp. 355–448.

STOUDER, R. E. (1941) Geology of the Big Clifty quadrangle: Kentucky Dept. Mines and Minerals, Geol. Div. Bull. ser. 8, no. 7, pp. 1–72.

TIPPIE, F. E. (1943) Subsurface stratigraphic sections near type Chester localities in southwestern Illinois: Illinois Acad. Sci. Trans., vol. 35, pp. 141–144; Illinois Geol. Survey Circ. 91.

——— (1944) Insoluble residues of the Levias and Renault formations in Hardin County, Illinois: Illinois Acad. Sci. Trans., vol. 36, pp. 155–157; Illinois Geol. Survey Circ. 102.

——— (1945) Rosiclare-Fredonia contact in and adjacent to Hardin and Pope counties, Illinois: Am. Assoc. Petroleum Geologists Bull. vol. 29, pp. 1654–1663; Illinois Geol. Survey, Rept. Inv. 112.

ULRICH, E. O. (1917) The formations of the Chester series in western Kentucky and their correlatives elsewhere: Kentucky Geol. Survey, ser. 5, vol. 1, pt. 2, pp. 1–272.

WELLER, J. M., and SUTTON, A. H. (1940) Mississippian border of Eastern Interior basin: Am. Assoc. Petroleum Geologists Bull. 24, pp. 765–858; Illinois Geol. Survey, Rept. Inv. 62.

WELLER, STUART (1913) Stratigraphy of the Chester group in southwestern Illinois: Illinois Acad. Sci. Trans. vol. 6, pp. 118–129.

——— (1920*a*) The Chester series in Illinois: Jour. Geology, vol. 28, pp. 281–303, 395–416.

——— (1920*b*) The geology of Hardin County and the adjoining part of Pope County: Illinois Geol. Survey Bull. 41.

WORKMAN, L. E. (1940) Subsurface geology of the Chester series in Illinois: Am. Assoc. Petroleum Geologists Bull. vol. 24, pp. 209–224; Illinois Geol. Survey Rept. Inv. 61.

# OSAGE-MERAMEC CONTACT[1]

L. R. LAUDON
University of Kansas, Lawrence, Kansas

ABSTRACT

Evidence is advanced that indicates a major break within the Mississippian system of rocks at the end of the Osage epoch. A marked physical break is demonstrated in the upper Mississippi Valley area, the Batesville, Arkansas area, the northeastern Oklahoma area, the subsurface of Kansas, the southwestern New Mexico area, the northern Montana area, the Banff area in British Columbia, and the Wapiti Lake area in northern British Columbia. Widespread differences in the distribution of Osage and Meramec rocks are indicated, and, finally, attention is called to one of the most remarkable faunal breaks in the entire Paleozoic era.

## INTRODUCTION

Serious divergence of opinion regarding recognition and usage of series and group units within the Mississippian system has existed among Mississippian stratigraphic workers for a considerable time. Evidence has accumulated that appears to indicate a major break at the end of the Osage epoch. This break appears to correspond with the major break recorded between rocks of Tournaisian and Viséan age in the European section. The evidence indicates the desirability of retaining the standard terms "Kinderhook," "Osage," "Meramec," and "Chester" and militates against any usage whereby Osage and Meramec rocks are classed together as a unit.

## PREVIOUS WORK

The name "Osage" was proposed by H. S. Williams (1891, p. 169) for rocks of supposed Burlington and Keokuk age exposed along Osage River near Osceola, Missouri. It has since been determined (Kaiser, 1945, p. 34) that no rocks of Keokuk age are exposed in the area and that most of the Upper Burlington is missing. Therefore, it can be seen that the type section was unfortunately located.

The Osage group has since been expanded by various workers to include several early Osage formations. The New Providence formation of Kentucky and Tennessee (Moore, 1928, p. 167; Stockdale, 1939, p. 49), the Fern Glen formation of Missouri (Ulrich, 1911, pl. 29; S. Weller and St. Clair, 1928, p. 166; J. M. Weller and Sutton, 1940, p. 793; Cline, 1934, p. 1136; Stockdale, 1939, p. 41), the St. Joe formation of Arkansas and Oklahoma (Moore, 1928, p. 167; Laudon, 1939, p. 325), and the Lake Valley formation of New Mexico (Laudon and Bowsher, 1941, p. 2133) are all now classed as early Osage in age.

The only serious divergence of opinion with regard to the lower contact of the Osage is voiced by Branson (1938, p. 8), who believes that the Chouteau limestone and Fern Glen formation of Missouri are of the same age, and therefore Branson places the Fern Glen formation in the Kinderhook group.

The problem concerning the upper contact of the Osage group has been open to question for some time. The Warsaw formation (Hall, 1857) is considered by many workers (Van Tuyl, 1922, p. 184; Laudon, 1931, p. 341; Stockdale, 1939, p. 16) to be of Osage age and to contain a late Osage fauna. In general, many workers east of the Mississippi River have classed the Warsaw formation as basal

[1] Manuscript received December 26, 1947.

Meramec in age (Ulrich, 1911, pl. 29; J. M. Weller and Sutton, 1940, p. 810).

The name "Meramec" was proposed by E. O. Ulrich (1904, p. 110) for the Warsaw, Spergen, and St. Louis formations. The Ste. Genevieve limestone, now commonly classed with the Meramec group, was then considered as basal Chester in age.

The name "Valmeyer" was suggested by Weller and Sutton for all rocks of Osage and Meramec age and was first used in classification by Moore (1933, p. 262).

Problems concerning the Osage-Meramec contact cannot be solved at the type sections of either the Osage or the Meramec series because both sections are incomplete. At the type section of the Osage, the Upper Burlington, Keokuk, Salem, and St. Louis formations are all missing. The Upper Burlington and Keokuk beds are exposed at other places in west-central Missouri, the Warsaw and Salem formations are not represented at all, and the St. Louis formation is known from only one small area. At the type section of the Meramec series, beds of Osage age are not exposed. In the vicinity, beds of Keokuk, Burlington, Reeds Spring, and Fern Glen age are exposed, but the Warsaw formation is not represented.

### THE WARSAW PROBLEM

Because the Warsaw formation has been classified either with beds of Meramec age or with beds of Osage age, it appears logical to present the evidence that is available at the type section of the Warsaw formation in Soap Factory Hollow near Warsaw, Illinois. Re-examination of the type section and numerous other sections on both the Illinois and the Iowa sides of the Mississippi River has yielded considerable information.

At the type section, slightly over 118 feet of strata occur between the last crystalline crinoidal beds of the Keokuk formation and the contact of the St. Louis limestone. No beds of Salem age were recognized in this area. No evidence was found at the type section for separating some 10 feet of cross-bedded calcareous sandstone, that lies immediately beneath the St. Louis limestone in the area, from the Warsaw formation. This sandstone has been considered to represent the Salem formation (S. Weller, 1908, p. 163).

No evidence of disconformity was found between rocks of Keokuk age and the Warsaw beds at the type section or at several other sections studied in the area. The Keokuk beds become more shaly in the upper portion and grade insensibly up into the Warsaw formation. Several measurements from the top of the crystalline crinoidal Keokuk beds up to known horizons within the Warsaw were made in an attempt to demonstrate transgressive overlap against the Keokuk surface. In all cases the thickness was very constant, and no evidence of disconformity was noted.

No evidence of a physical break was found within the Warsaw formation. A study of various intervals within the Warsaw formation on several sections demonstrated no evidence of unconformity within the formation. Abundant faunas do appear quite suddenly in the middle part of the section, but in most cases the species are identical with forms that occur in the shaly beds in the upper portion of the Keokuk formation.

The contact between the Warsaw formation and the overlying St. Louis limestone is remarkably unconformable. Deep channeling and considerable relief can be demonstrated on the Warsaw surface within very short distances. True

conglomerates (not St. Louis brecciated limestone) are present in some of the deeper channels on the Warsaw surface. At several exposures large blocks of brown, upper Warsaw dolomite beds are re-worked into the basal St. Louis beds.

Because of the general confused state of the Warsaw problem we have deemed it wise to disregard completely all previously published lists of faunas. Accordingly, faunas were collected from both Warsaw and Keokuk formations in the type areas. Our new collections are not large, but the common forms of both formations are represented. Analysis of the Warsaw fauna indicates very close relationships with that of the underlying Keokuk formation. Thirty species of abundant Warsaw forms were identified, and twenty-seven of them were found in the underlying Keokuk rocks. The occurrence of *Metablastus wortheni*, *Dizygocrinus indianensis*, *Macrocrinus jucundus*, *Spirifer tenuicostatum*, *Cleiothyrindina obmaxima*, *Echinoconchus alternatus*, *Athyris lamellosa*, *Reticularia pseudolineata*, *Lioclema gracillamum*, *Hemitrypa proutana*, and *Bactropora simplex* in the Warsaw fauna indicates very strong affinities with the underlying Keokuk beds.

Comparison of the Warsaw fauna with twenty abundant species from the type area of the Salem formation shows almost no affinities at all. Only one species, *Pentremites conoideus*, was found in both faunas.

This observation contrasts sharply with general opinion regarding the relationships of the Warsaw-Salem faunas (J. M. Weller and Sutton, 1940, p. 808). Because of the relatively small portion of the Salem fauna that was used in our observations, it is entirely possible that we have arrived at an erroneous conclusion. Our Salem fauna consisted entirely of the diminutive forms from the oölite facies of the Salem formation, and because this is a facies fauna it may be entirely misleading. The writer has not had opportunity to study the field relationships of equivalent formations in southern Indiana.

It seems probable that the Warsaw beds exposed in the type area are the last deposits of the Osage seas. The calcareous facies of the Keokuk and older Osage formations in general become more clastic eastward toward the Appalachian Mountains (Stockdale, 1939, pp. 16, 18). It is normal that, as uplift progressed in eastern areas near the end of the Osage epoch, the clastic facies shifted farther to the west, away from the mountains that were forming. The Warsaw does resemble the clastic facies of the older Osage formations farther east and does not resemble typical Meramec rocks. The Salem limestone is considered in this paper as the base of the Meramec series in the upper Mississippi Valley area, but our evidence for placing it in the base of the Meramec is weak and inconclusive and needs further study. If the Salem limestone is a calcareous facies development within the Warsaw formation, as suggested by Weller and Sutton (1940, p. 803), this classification will need serious revision.

A spectacular unconformity, as well as distinctive faunal and lithologic breaks, marks the contact between Meramec and older Mississippian rocks in all areas in which the writer has had experience in western North America. A few areas have been chosen to illustrate this physical break.

### BATESVILLE AREA, ARKANSAS

Confusion in the Batesville area (fig. 1) is due mainly to failure of workers to recognize different zones within the Osage section (Girty, 1915, p. 5). The

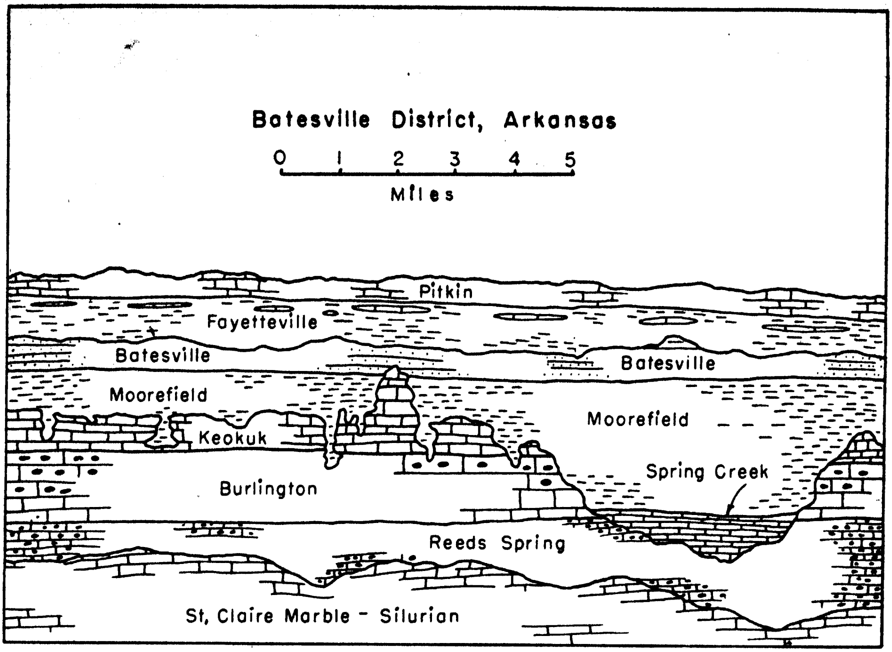



FIG. 1.—Diagrammatic sketch, showing the general stratigraphic relationships of Mississippian formations in the Batesville area, Arkansas.

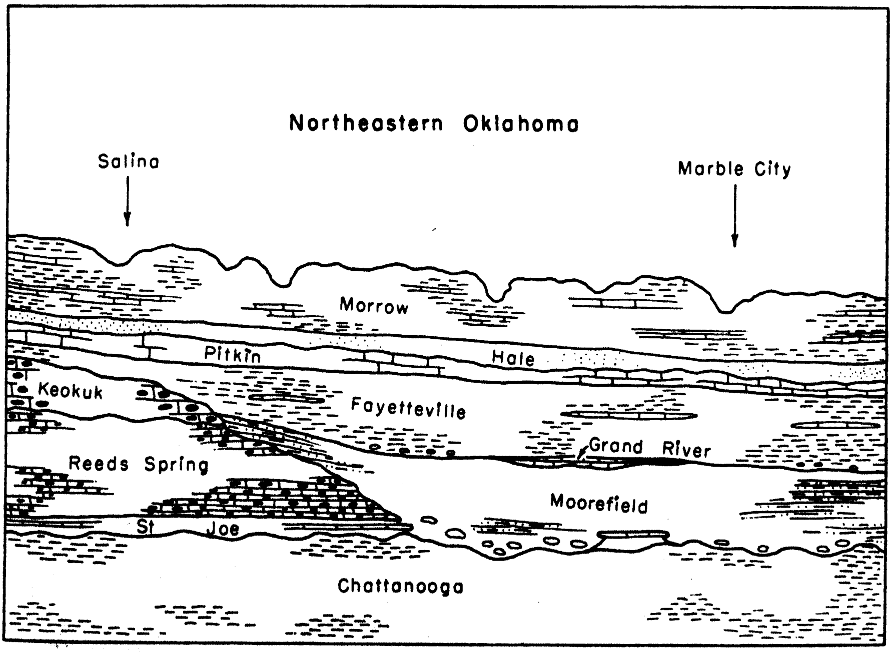



FIG. 2.—Diagrammatic sketch, showing the relationships of Osage rocks to the younger Mississippian rocks in northeastern Oklahoma.

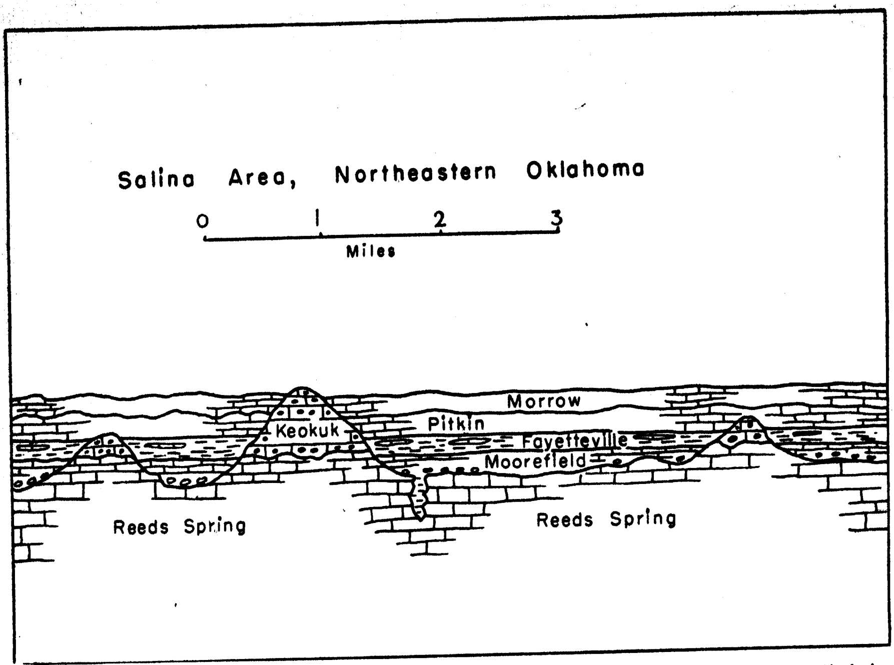



FIG. 3.—Diagrammatic sketch, showing buried hills of Osage rock rising through younger Mississippian strata in the Salina district, Oklahoma.

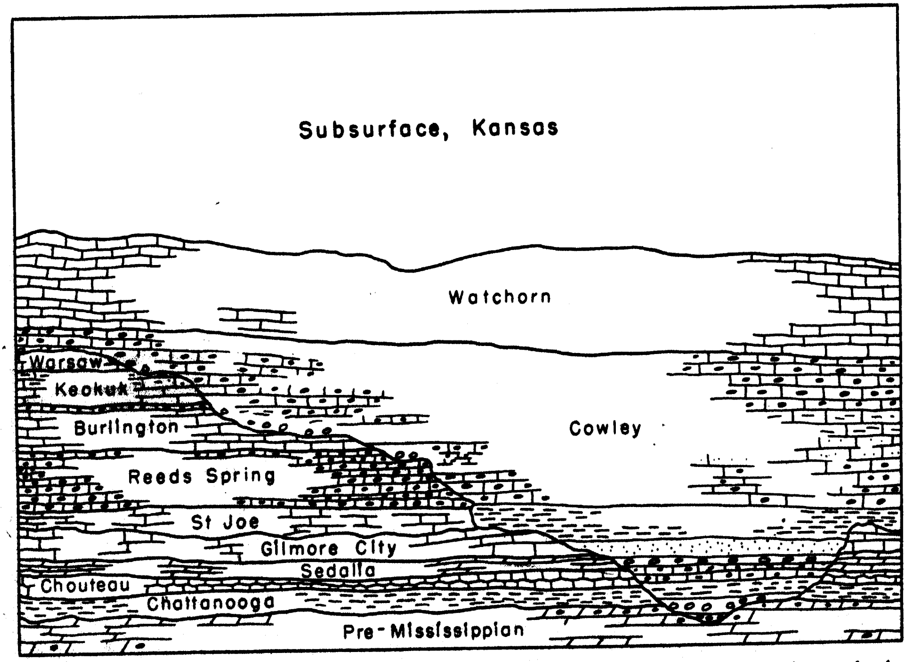



FIG. 4.—Diagrammatic sketch, showing the subsurface relationships of Mississippian rocks in Kansas.

St. Joe limestone is absent over most of the area, and the Reeds Spring cherty limestone rests on the eroded surface of the St. Claire limestone of Silurian age. Typical Burlington limestone is present, but apparently only the upper Burlington beds are represented. The Keokuk is represented by the typical southern and southwestern Ozark facies in which massive, gray, crinoidal, noncherty limestone beds are developed. It is possible that the upper portion of these limestone beds actually is a southwestward equivalent of the Warsaw formation, although what meager fauna is present belongs typically with the Keokuk.

The Moorefield and Batesville formations of early Meramec age are found in contact with beds of the Osage section ranging in age from Reeds Spring to late Keokuk within a few miles of Batesville. A veritable karst topography was developed on the Osage surface before the deposition of Meramec sediments. Deep sinkholes penetrate the Osage rocks and in some of the old quarries east of Batesville extend over half the quarry face. At Ruddells Mill, west of Batesville, the Spring Creek member of the Moorefield shale lies in contact with the Reeds Spring formation. The Spring Creek sediments were deposited only in the lowest erosional channels on the Osage surface, explaining its very local distribution in the area. The total relief on this surface is well over 150 feet within very short distances.

###### NORTHEASTERN OKLAHOMA AREA

The entire Osage section is truncated and overlapped by rocks of Meramec age from south to north in northeastern Oklahoma (fig. 2). First recognition of this truncation was presented by Cline (1934, p. 1157). Just north of Marble City the Moorefield formation is locally in contact with the Chattanooga shale and a conglomerate consisting of reworked Osage chert is present in the base of the Moorefield.

In the Salina district (fig. 3) in northeastern Oklahoma remnant hills of Reeds Spring and Keokuk chert extend through the thin northern edge of Moorefield, Chester, and Morrow formations. Some of these chert hills projected high enough to be overlapped by Cherokee sediments. They are rather conspicuous topographic features in the area, since the soft Cherokee sediments have been removed by erosion, exposing the Morrow surface.

###### SUBSURFACE RELATIONS IN KANSAS

The complex subsurface relationships of the Mississippian rocks in Kansas (fig. 4) have been studied in detail by Lee (1940, p. 66). Remarkable relief was developed on the Osage surface before the deposition of the Cowley sediments. The Cowley formation as defined by Lee is roughly equivalent to the Moorefield formation of the southern Ozarks region. In some places in western Kansas the Cowley formation rests directly on pre-Mississippian rocks.

###### NORTH-CENTRAL IOWA AREA

No Osage rocks are exposed in north-central Iowa, but the remarkably rough Kinderhook (Gilmore City) surface (fig. 5) is probably a result of post-Osage erosion. In the Gilmore City area remarkable sinkholes have been formed in the Gilmore City limestone (Laudon, 1933, p. 21). These sinkholes, although in most cases filled with sediments of St. Louis age, have been reopened by recent subsurface drainage channels, so that the farmers of the area now drain the surface water from their farms into them. In the quarries at Gilmore City considerable portions of the limestone have been

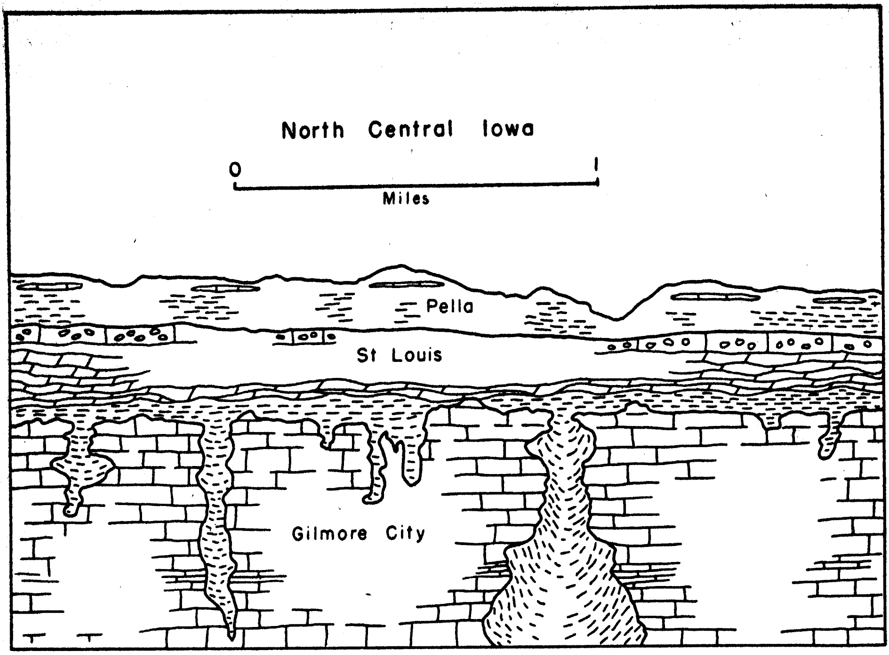



FIG. 5.—Diagrammatic sketch, showing the remarkable sinkholes developed on the Gilmore City limestone and later filled with sediments of St. Louis age.

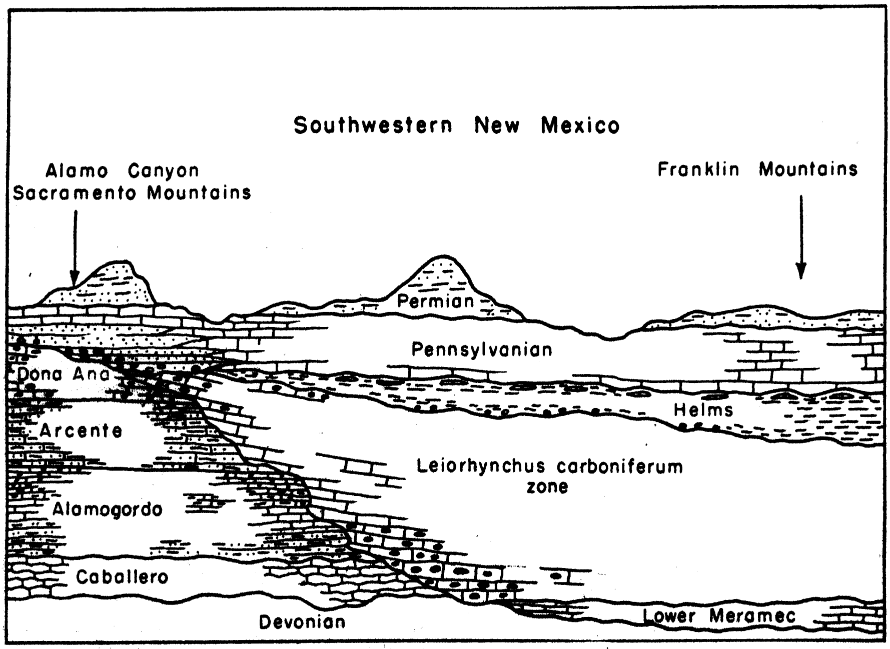



FIG. 6.—Diagrammatic sketch, showing the relationships of Kinderhook and Osage sediments to sediments of early Meramec age in southwestern New Mexico.

rendered almost unworkable because of channels filled with shales of St. Louis age.

#### SOUTHWESTERN NEW MEXICO AREA

The entire Kinderhook (Caballero) and Osage (Lake Valley) section is truncated southward in the Sacramento and San Andres Mountain areas in New Mexico and overlapped by sediments of early Meramec age (fig. 6). Siltstones carrying the *Leiorhynchus carboniferum* fauna, correlatives of the Moorefield formation of the Ozark area, are excellently exposed in the Hueco and Franklin mountain areas in New Mexico, where they rest on Devonian sediments. These early Meramec sediments transgressively overlap northward onto Kinderhook and Osage sediments in the Sacramento and San Andres Mountains and in their northernmost exposures may be seen in contact with the upper portion of the Lake Valley formation.

#### GALLATIN BASIN AREA, MONTANA

The Madison formation of Montana is usually divided into two members, Lodgepole below and Mission Canyon above (Sloss and Hamblin, 1942, p. 313). Recent work by Leonard (1946, p. 25) has demonstrated a widespread physical break within the Mission Canyon member. In the Gallatin Basin area (fig. 7) massive gray limestone beds carrying a *Lithostrotion* fauna rest with marked unconformity on the lower limestone beds of the Mission Canyon member. Faunas collected in the lower portion of the Mission Canyon indicate late Kinderhook age. No beds of Osage age were found in the area of the type section of the Madison limestone.

Deep sinkhole-like channels have been carved into the restricted Mission Canyon member, and re-worked materials of the underlying Mission Canyon limestone have been deposited in these channels. Excellent exposures showing these relationships have been developed in the north canyon wall of Yellowstone River, a few miles south of Livingston, Montana.

#### BANFF AREA, BRITISH COLUMBIA

In the Banff area in British Columbia confusion has arisen concerning the Mississippian stratigraphy of the area due mainly to erroneous identification of fossils and to the failure to recognize an unconformity that exists within the Rundle formation. The Mississippian section at Banff in British Columbia (fig. 8) consists of a shaly limestone series at the base that carries a Kinderhook fauna and is a direct correlative of the Lodgepole of Montana. This shaly series is overlain by the massive gray limestone beds of the Rundle formation. The massive limestone beds of the lower portion of the Rundle formation carry typical late Kinderhook faunas and correlate with the restricted Mission Canyon section in Montana.

The upper portion of the Rundle formation consists of massive gray, cherty limestone that carries a *Lithostrotion* fauna and is the equivalent, at least in part, of the St. Louis limestone of the Mississippi Valley. It is quite certain, however, that the Meramec portion of the Rundle formation represents considerably more time than does the St. Louis limestone of the Mississippi Valley.

The Meramec portion of the Rundle formation rests with marked unconformity on limestones of late Kinderhook age in the Banff area. The contact is everywhere overlain by a basal shaly zone that often carries phosphatic concreations and fish teeth. The change in

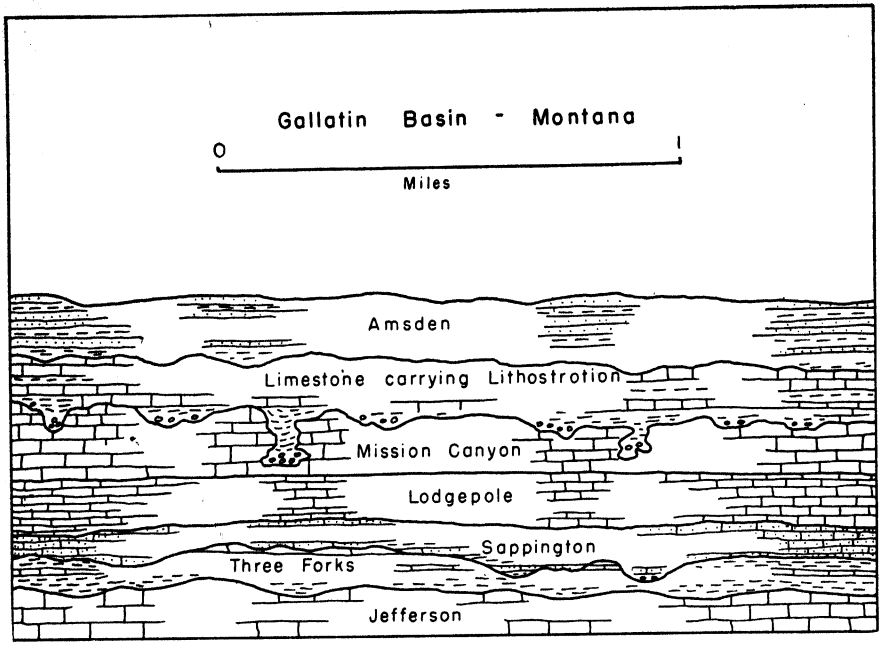



FIG. 7.—Diagrammatic sketch, showing the relationship of Meramec sediments to the underlying Kinderhook sediments in the Gallatin Basin area in northern Montana.

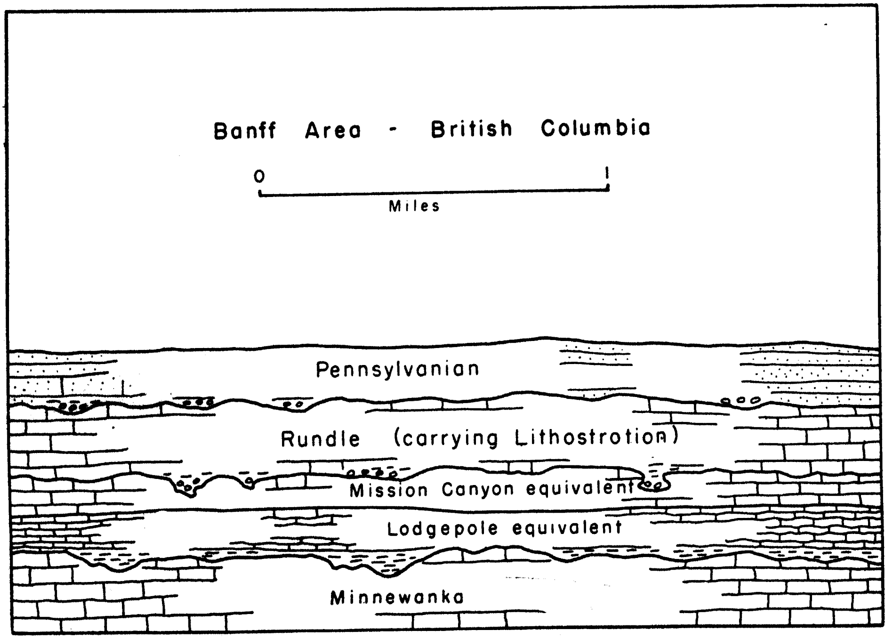



**FIG. 8.—Diagrammatic sketch, showing the relationship of Kinderhook beds to the overlying Meramec beds in the Banff area, British Columbia.**

lithology is abrupt, although both are limestones. Typical St. Louis lithology appears immediately above this contact. Sufficient work was not completed in the area to determine the amount of relief of the Kinderhook surface.

lain by rhythmic sequences of limestone beds similar in lithology and fauna to the Lodgepole formation of Montana. These rhythmic limestones are overlain by a thick accumulation of gray, massive limestone beds, approximately equiva-

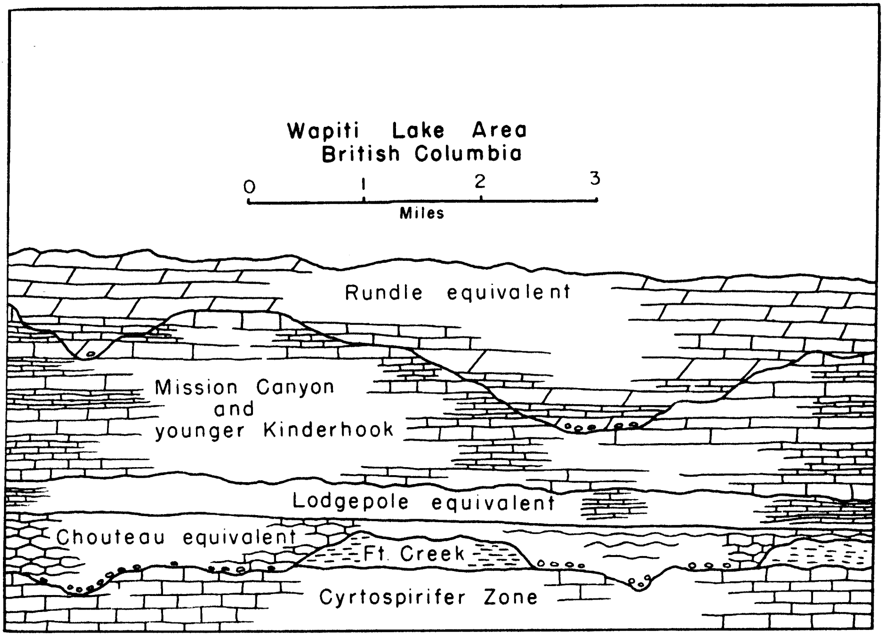



FIG. 9.—Diagrammatic sketch, showing the relationships of rocks of Meramec age to the underlying Kinderhook rocks in the Wapiti Lake area, British Columbia.

lent in age to the lower portion of the Mission Canyon formation of Montana. No rocks of Osage age were recognized in the area. The Kinderhook rocks are overlain by brown, sugary dolomites and gray, cherty limestone beds that carry a rich *Lithostrotion* fauna. These beds are similar to some in the upper portion of the Rundle formation of the Banff area. The beds of Meramec age rest on rocks of late Kinderhook age with marked unconformity. Within a distance of less than 5 miles they were found to overlap against considerably more than 500 feet of Kinderhook strata.

#### WAPITI LAKE AREA, BRITISH COLUMBIA

The Wapiti Lake area in British Columbia lies at the east edge of the Canadian Rocky Mountains, south of the Peace River, approximately 100 miles directly west of Grande Prairie, Alberta. Mississippian rocks (fig. 9) are unusually well displayed in the area and are very fossiliferous, making close correlation possible.

The basal beds consist of nodular, gray limestone carrying an abundant fauna, remarkably similar to that of the Chouteau formation of the Mississippi Valley area. The nodular beds are over-

### DISTRIBUTION OF MERAMEC VERSUS OSAGE ROCKS

If the distribution of Osage rocks is compared to the distribution of Meramec rocks, an important break between the two series is suggested.

Osage rocks have a rather local distribution in comparison with Meramec rocks (fig. 10). Osage rocks are known today from the Appalachian slope region, the Mississippi Valley, and Mid-Continent regions. In the west they are confined to New Mexico, to parts of Arizona, and to local exposures in southern Nevada. They are entirely missing in the northern Rockies and in the Canadian Rockies.

In contrast to this, rocks of Meramec age widely overlap onto Kinderhook and older surfaces (fig. 11). In western United States the distribution of Meramec, however, extends into the northern Rockies, the Canadian Rockies, and Alaska.

### PALEONTOLOGY

Careful analysis of the ranges of fossils within Mississippian rocks reveals one of the most remarkable faunal breaks in the whole Paleozoic sequence at the end of the Osage series. Kinderhook and Osage species appear to be closely related and often exhibit very complete evolutionary series. Meramec and Chester species show very little relationship to those of the Osage and, in general, are much more closely allied with the younger Pennsylvanian species than with the older Mississippian forms.

Most remarkable is the almost com-

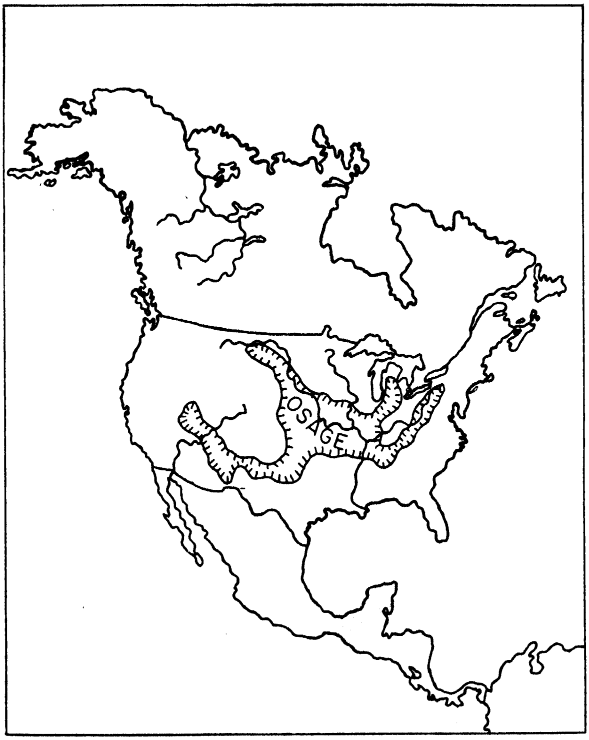
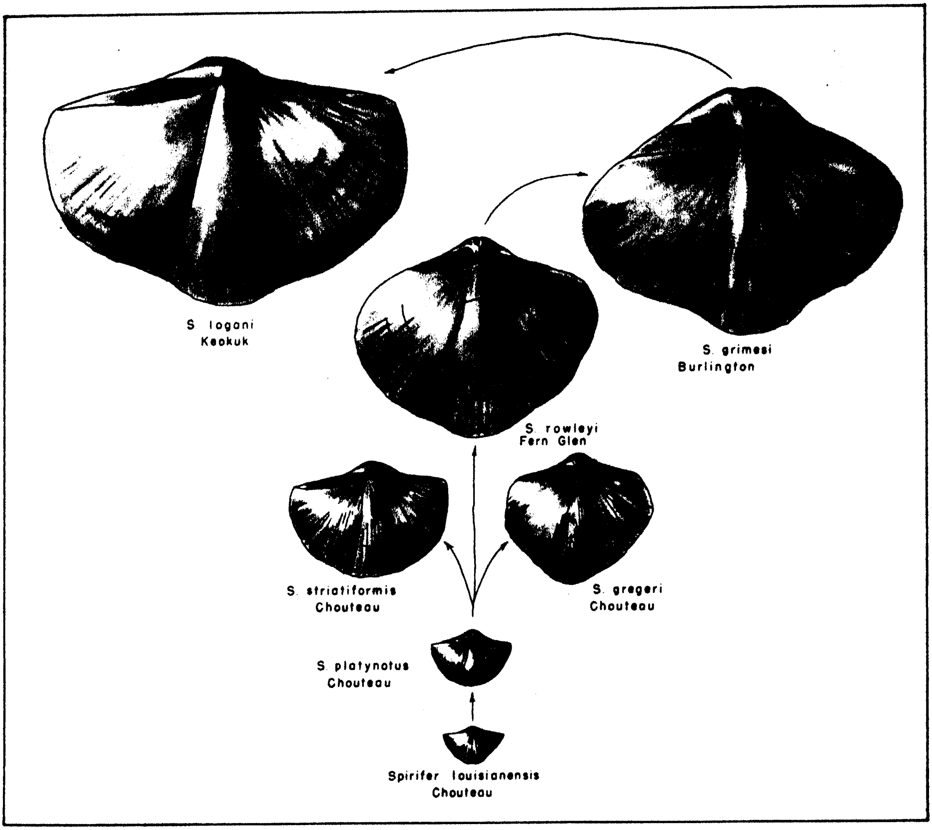
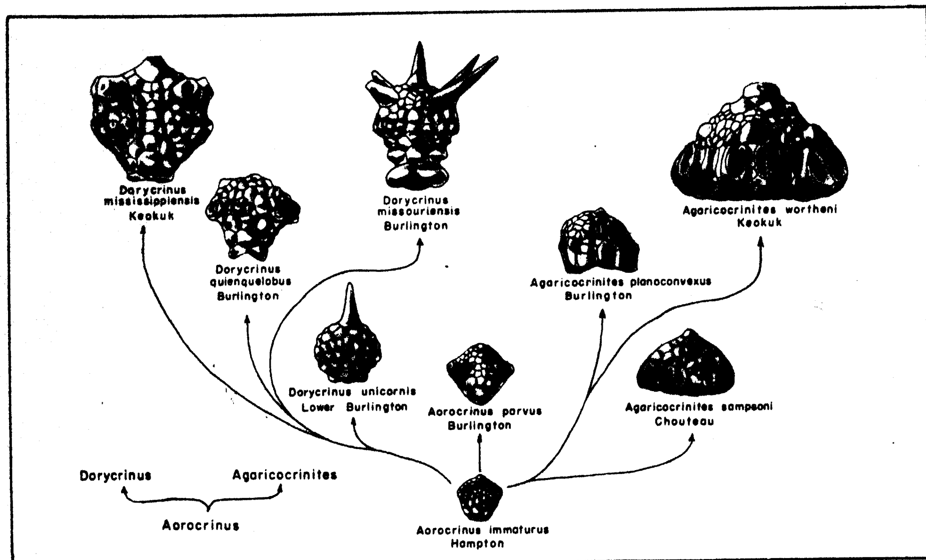
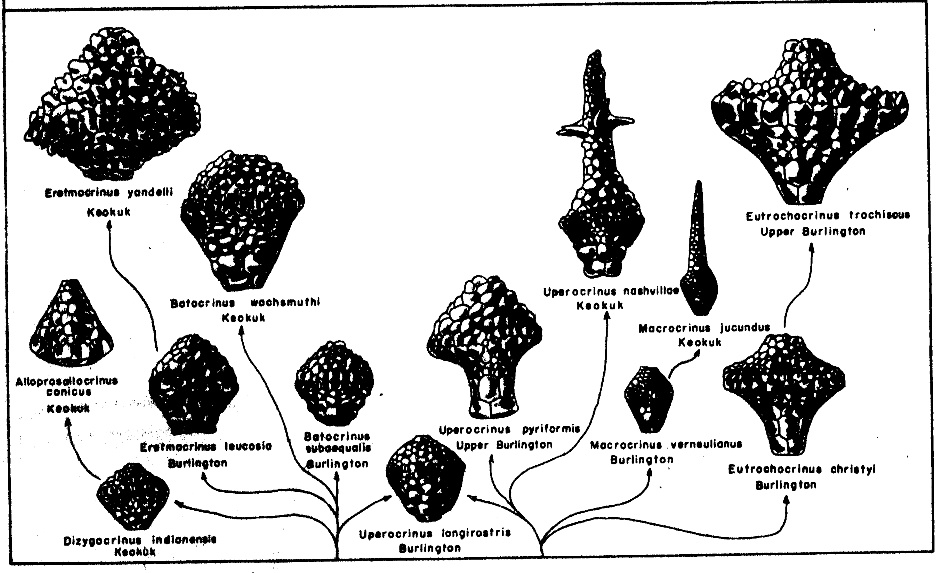



FIG. 10.—Map showing the general distribution of Osage rocks in North America



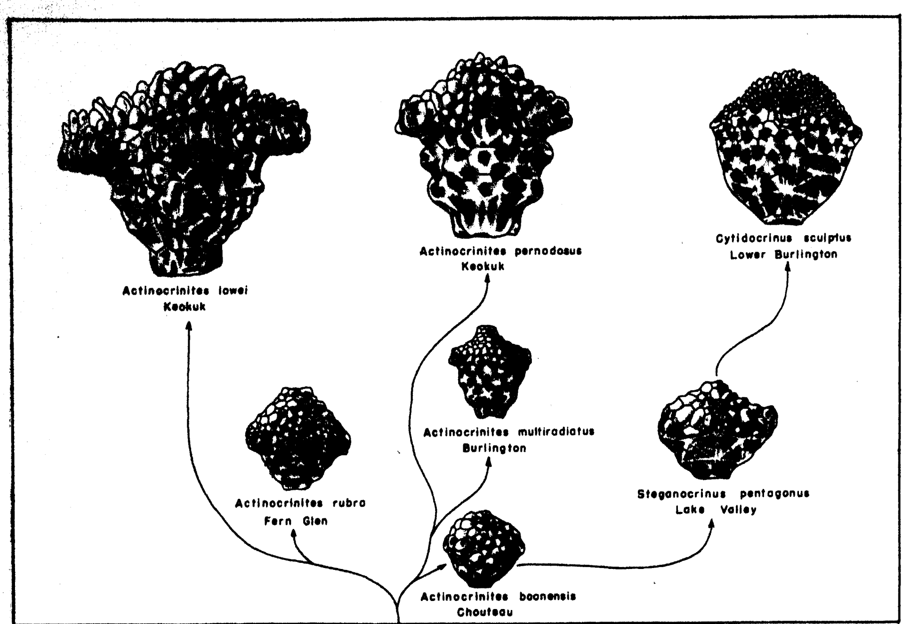

Evolution chart, showing the Kinderhook and Osage species of the genus *Spirifer*. None of these species occur in Meramec or Chester rocks.

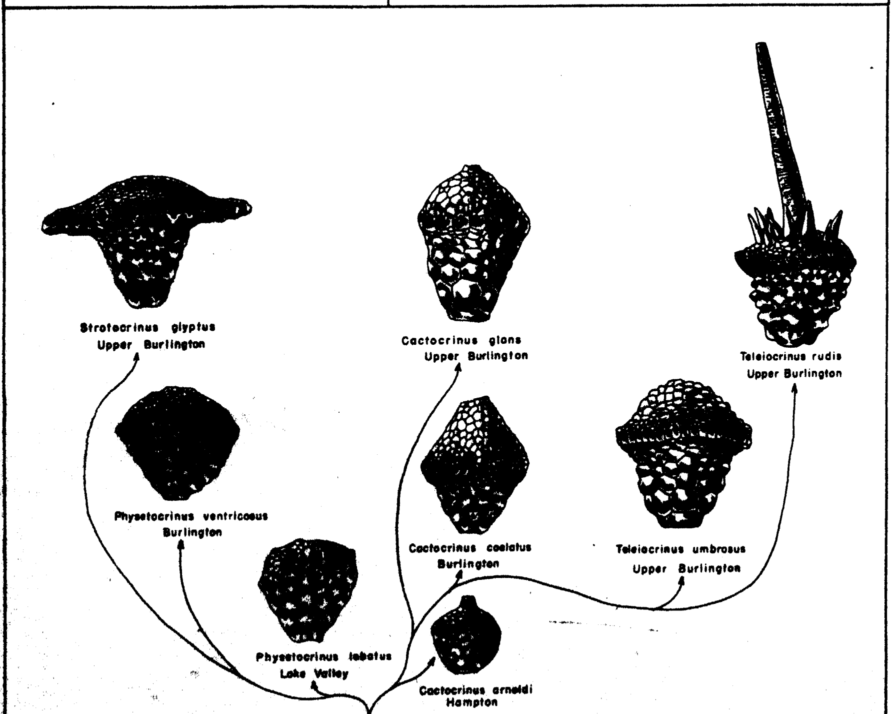



*Upper*, Evolution chart, showing Kinderhook and Osage species in the family Desmidocrinidae. None of these genera or species occur in Meramec or Chester rocks.

*Lower*, Evolution chart, showing Kinderhook and Osage species in the family Batocrinidae. None of these genera or species occur in Meramec or Chester rocks.


Actinocrinites lowei
Keokuk
Actinocrinites pernodosus
Keokuk
Cytidocrinus sculptus
Lower Burlington
Actinocrinites multiradiatus
Burlington
Actinocrinites rubra
Fern Glen
Steganocrinus pentagonus
Lake Valley
Actinocrinites boonensis
Chouteau


Strotocrinus glyptus
Upper Burlington
Cactocrinus glans
Upper Burlington
Teleiocrinus rudis
Upper Burlington
Physetocrinus ventricosus
Burlington
Cactocrinus coelatus
Burlington
Teleiocrinus umbrosus
Upper Burlington
Physetocrinus lobatus
Lake Valley
Cactocrinus arnoldi
Hampton

plete annihilation of the highly specialized Osage fauna. Of the 54 most commonly occurring Kinderhook and Osage brachiopod genera, 26 are limited to the Kinderhook and Osage (pl. 1). Only 1 genus is limited to the Meramec and Chester. Only 1 known species (*Linoproductus ovatus*) crosses the Osage-Meramec line. The most outstanding faunal change is the remarkable disappearance of the camerate crinoids (fig. 12). Of 321 abundant Mississippian camerate crinoid species, 298 are restricted to Kinderhook and Osage rocks, and only 23 occur in rocks of Meramec and Chester age. Of 20 species in the family Rhodocrinitidae, none crosses the Osage-Meramec line. Of 127 species falling in the Desmidocrinidae and Bactocrinidae, none crosses the Osage-Meramec line (pl. 2). Of 66 commonly occurring species in the Actinocrinitidae (pl. 3), none crosses the Osage-Meramec line. Of 58 species in the Platycrinitidae, 57 fail to cross the Osage-Meramec line, and only 1 known species is restricted to the Upper Mississippian rocks. Of the 46 commonly occurring species of the Dicho-

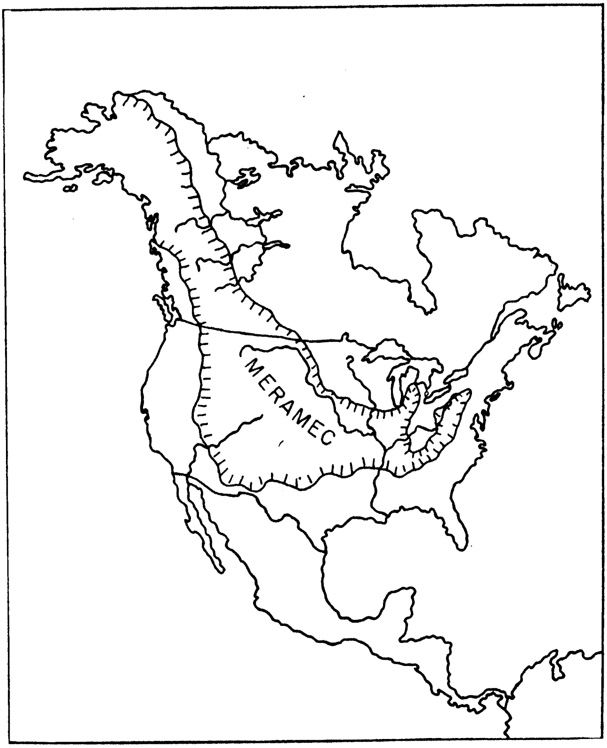



FIG. 11.—Map showing the general distribution of Meramec rocks in North America

PLATE 3

*Upper*, evolution chart showing Kinderhook and Osage species of *Actinocrinites*, *Steganocrinus*, and *Cytidocrinus*. None of these genera or species occur in Meramec or Chester rocks.

*Lower*, Evolution chart showing Kinderhook and Osage species of *Cactocrinus*, *Physetocrinus*, *Strotocrinus*, and *Teleiocrinus*. None of these genera and species occur in Meramec or Chester rocks.

crinidae, 19 are found in the Upper Mississippian rocks, but all are easily distinguishable from Kinderhook and Osage forms. No known species of *Dichocrinus* crosses the Osage-Meramec line.

Of 57 commonly occurring Mississippian species of flexible crinoids, only 7 of 37 commonly occurring Mississippian inadunate crinoid genera, 23 do not cross the Osage-Meramec line. Four genera—*Graphiocrinus*, *Linocrinus*, *Culmicrinus*, and *Cyathocrinus*—cross the Osage-Meramec line, but no species are known to do so. Two genera of inadunate

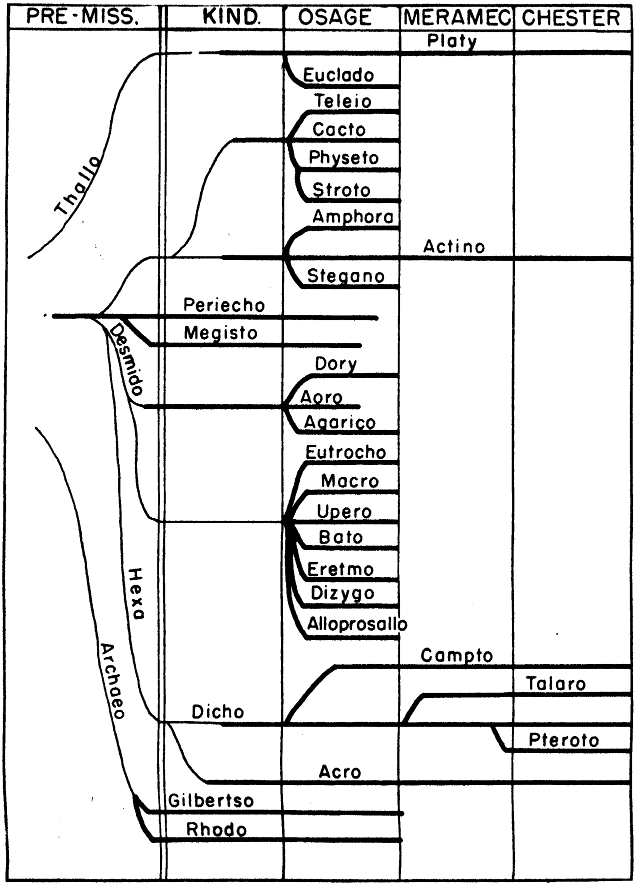



FIG. 12.—Chart showing the range of Mississippian camerate crinoid genera

are known from Meramec and Chester rocks. None of the species in the Lecanocrinidae and Sagenocrinidae and only 1 of the species in the Ichthyocrinidae occur in the Upper Mississippian rocks. Six of the 28 species of the Taxocrinidae are found in the Upper Mississippian rocks, but no species crosses the line. Of crinoids start in the Meramec, and 8 start in the Chester.

## TOURNAISIAN-VISÉAN BREAK IN THE BRITISH ISLES

Throughout most of the more definitely offshore areas of Mississippian rocks, namely, the Mississippi Valley, the Mid-

Continent, and western North America, the lower Kinderhook beds consist of clastic shales, silts, and calcareous mudstones with intercalated nodular limestone beds. Beds similar in lithology and remarkably similar in fauna occupy the more definitely marine portion of the Tournaisian section in the British Isles. The zones of Km, $K_1$, and $K_2$ correlate, in general, with our Chattanooga, Saverton, Louisiana, Hannibal, Chouteau, Sedalia, Caballero, Lodgepole, and upper Banff limestone sections. Exact correlations cannot be made, but faunas and lithology correspond remarkably closely.

Massive gray limestone beds, many of them oölitic, interbedded with darker, more clastic, shaly limestone beds usually follow next in stratigraphic order in North America. These beds, except for the Gilmore City limestone of Iowa, are not present in the upper Mississippi Valley province or in the Mid-Continent region. However, they are excellently displayed throughout most of western North America. The massive gray limestone beds of the Leadville limestone of Colorado, most of the Pahasapa limestone of the Black Hills, the basal portion of the Rundle limestone in southern British Columbia, and a thick series of limestone beds in the Wapiti Lake area in northern British Columbia belong to this period of deposition. In western North America these beds are characterized by the abundance of Zaphrentid corals and late Kinderhook crinoid faunas. The massive, gray, crinoidal limestone beds of the $Z_1$ and $Z_2$ zones of the British Isles correlate, in general, with this portion of the American section. The abundant occurrence of Zaphrentid corals and a remarkably similar brachiopod fauna in the British section support this correlation.

The deposition of the gray, massive limestone beds in the late Kinderhook marks one of the maximum floods of Mississippian seas in North America. Immediately following this flood, the seas began to withdraw rapidly from North America, resulting in considerably less widespread deposition of Osage sediments. Contrary to popular opinion, no beds of Osage age are known in British Columbia, northern Montana, the Black Hills, and Colorado. Beds of Osage age are limited to the Mississippi, Ohio, and Tennessee drainage areas, the Mid-Continent region, and local parts of New Mexico, Arizona, and Nevada. The Osage beds, in general, represent the final deposition before the important Osage-Meramec break. Such highly specialized evolutionary types as characterize the late Osage beds commonly occur during periods of rapid environmental change. *Caninia* first occurs in abundance in the American section in beds of early Osage age. The massive crinoidal limestone beds of the $C_1$ zone of the British section correlate very closely with the Osage beds of America. The first abundant occurrence of *Caninia* occurs in these beds in the British section, and, as in America, highly developed crinoidal bioherms are developed in the Coplow Knoll area. Such common Osage crinoid genera as *Actinocrinites* and *Gilbertsocrinus* occur abundantly in the Coplow Knoll fauna. These genera do not range beyond the Osage in North America.

If we neglect the highly problematical Warsaw and Salem formations of the Mississippi Valley area, the next widespread Mississippian deposition in North America is characterized by widespread accumulation of highly clastic siltstones and shales that are occasionally somewhat calcareous. All over North America they are marked by the occurrence of the *Leiorhynchus carboniferum* fauna. These beds are represented by the Moorefield and Batesville formations of the Ozark

province, the Cowley formation of the subsurface of Kansas, the Sycamore limestone of the Arbuckle Mountains of Oklahoma, the lower portion of the Helms formation of New Mexico, and the thick accumulations of sediments in northern British Columbia, in Yukon, and in the Calico Bluff area in Alaska. At no known place in North America are beds of Moorefield age actually in contact with younger beds of Meramec age on the surface, unless it is in the northern portion of Alaska in the Cape Lisburne area. Indicated stratigraphic position of the Moorefield formation and its equivalents has been determined entirely on the basis of the stage of evolution of the fauna, and the evidence is not conclusive. It is entirely possible that the fauna is of late Meramec age. The zone of $C_2$ in the British section corresponds in lithology and stratigraphic position with that of the *L. carboniferum* beds of North America. However, the fauna of the $C_2$ zone has very little in common with that of the *L. carboniferum* beds of North America. The zone of $C_2$ in the British Isles lies with marked unconformity on underlying strata and carries *Lithostrotion* and typical Meramec fossils. The Tournaisian-Viséan break is often placed between the zones of $C_1$ and $C_2$ by British workers. Thus it can be seen that the Tournaisian-Viséan break in the Lower Carboniferous section of the British Isles is directly comparable to the Osage-Meramec break in North America.

## REFERENCES CITED

BRANSON, E. B. (1938) Stratigraphy and paleontology of the Lower Mississippian of Missouri: Univ. Missouri Studies, vol. 13, no. 3, pp. 1–208.

CLINE, L. M. (1934) Osage formations of southern Ozark region, Missouri, Arkansas, and Oklahoma: Am. Assoc. Petroleum Geologists Bull. 18, pp. 1132–1159.

GIRTY, G. H. (1915) Fauna of the so-called Boone chert near Batesville, Arkansas: U.S. Geol. Survey Bull. 595, pp. 1–45.

HALL, J. (1857) Observations upon the Carboniferous limestones of the Mississippi Valley: Am. Jour. Sci., 2d ser., vol. 23, pp. 187–203.

KAISER, C. P. (1945) Stratigraphy of the Mississippian formations of the Osage River in western Missouri, Master's thesis, Univ. Kansas Library.

LAUDON, L. R. (1931) The stratigraphy of the Kinderhook series of Iowa: Iowa Geol. Survey, vol. 25, pp. 335–451.

——— (1933) The stratigraphy and paleontology of the Gilmore City formation of Iowa: Iowa Univ. Studies in Nat. History, vol. 15, pp. 1–74.

——— (1939) Stratigraphy of Osage subseries of northeastern Oklahoma: Am. Assoc. Petroleum Geologists Bull. 23, pp. 325–338.

——— and BOWSHER, A. L. (1941) Mississippian formations of Sacramento Mountains, New Mexico: Am. Assoc. Petroleum Geologists Bull. 25, pp. 2107–2160.

LEE, WALLACE (1940) Subsurface Mississippian rocks of Kansas: Kansas Geol. Survey Bull. 33, pp. 1–114.

LEONARD, J. R. (1946) Mississippian stratigraphy of the Gallatin Basin, Montana, Master's thesis, Univ. Kansas Library.

MOORE, R. C. (1928) Early Mississippian formations in Missouri: Missouri Bur. Geology and Mines, 2d ser., vol. 21, pp. 1–283.

——— (1933) Historical geology, New York, McGraw-Hill Book Co.

SLOSS, L. L., and HAMBLIN, R. H. (1942) Stratigraphy and insoluble residues of the Madison group of Montana: Am. Assoc. Petroleum Geologists Bull. 26, pp. 305–335.

STOCKDALE, P. B. (1939) Lower Mississippian rocks of the east-central interior: Geol. Soc. America Special Paper 22, pp. 1–248.

ULRICH, E. O. (1911) Revision of the Paleozoic systems: Geol. Soc. America Bull. 22, pp. 281–680.

VAN TUYL, F. M. (1922) The stratigraphy of the Mississippian formations of Iowa: Iowa Geol. Survey, vol. 30, pp. 39–374.

WELLER, J. M., and SUTTON, A. H. (1940) Mississippian border of eastern interior basin: Am. Assoc. Petroleum Geologists Bull. 24, pp. 765–858.

WELLER, STUART (1908) The Salem limestone: Illinois Geol. Survey Bull. 8, pp. 81–102.

——— and ST. CLAIR, S. (1928) Geology of Ste. Genevieve County, Missouri: Missouri Bur. Geology and Mines, vol. 22, 2d ser., pp. 1–352.

WILLIAMS, H. S. (1891) Correlation papers, Devonian and Carboniferous: U.S. Geol. Survey Bull. 80, pp. 1–279.

# PROBLEM OF THE "MAYES" IN OKLAHOMA[1]

ERWIN L. SELK

Stanolind Oil and Gas Company, Oklahoma City

ABSTRACT

The Mayes formation consists of those beds of Mississippian age underlain by the Boone (Osagian) and overlain by the Fayetteville (Chesterian) in the outcrops in Mayes County, Oklahoma, and is generally considered to be equivalent in age to the Moorefield and Batesville formations. The name was proposed by L. C. Snider in 1915 and correlated as lower Chester. In 1927, George S. Buchanan applied the name to the dark, argillaceous, silty limestone of the subsurface Mississippian and correlated the lower part at least with the Meramec series. In 1930, Ira H. Cram introduced convincing evidence that the subsurface "Mayes" was a facies of the Boone and was therefore of Osage age. Since that time the name has persisted, as has the confusion and controversy. Much new evidence, by deep drilling in western and northwestern Oklahoma, supports the general opinion of subsurface geologists that the "Mayes" of the subsurface is of Osage age and is not a correlative of the Mayes formation of the outcrop section in northeastern Oklahoma.

## INTRODUCTION

"Mayes" is the name commonly applied to the clastic limestone of the subsurface Mississippian found between the Arbuckle Mountains and the Mississippian outcrops of northeastern Oklahoma. In the greater part of this area it overlies the Woodford shale of Kinderhook age and is overlain by the Mississippian black shales, herein referred to as "Caney."

A number of papers have been published in which the "Mayes" has been correlated with strata in the classic Mississippi River section. In most published reports it is classified as Meramecian. In one early publication correlation with the Osagian is suggested.

Subsurface geologists, who have traced the "Mayes" throughout the area in which it occurs, almost unanimously correlate it with the Osage series. According to this view, the "Mayes" is a remarkable change in facies, which is not uncommon in Mississippian rocks of Osage age.

The subsurface "Mayes" is correlated with the formation of that name in the outcrop section of northeastern Oklahoma by the authors of several published reports. These efforts to establish the age of the "Mayes" as Meramecian have not been convincing to the subsurface geologist acquainted with the problem. On the other hand, the only published assertion of Osage age of the "Mayes" has not been generally accepted.

Another approach to the age problem is provided by the subsurface section of the Mississippian found in the Anadarko Basin of southwestern Kansas and northwestern Oklahoma, where the thickest section of Mississippian deposits in this province is found. The correlation of these beds indicates the Osage age of the subsurface "Mayes" and precludes its correlation with the outcrop formation of that name.

The subdivisions of the Mississippian section other than those considered to be Osagian, are discussed in this paper only in so far as they pertain and contribute to the solution of the "Mayes" problem.

## EARLIER WORK ON MISSISSIPPIAN STRATIGRAPHY OF OKLAHOMA

The outcrops of the Mississippian strata in northeastern Oklahoma have been studied repeatedly since the turn of the century. Equivalents of the four major subdivisions of the classic Mississippi River section (table 1) were recognized early in the unraveling of the stratigraphy of this area. Later detailed work led to the naming of the individual members, formations, and groups and their probable correlation with the out-

[1] Manuscript received December 22, 1947.

crops of the adjacent southwest Ozark region of southwestern Missouri and northwestern Arkansas.

Some of the early work was done by L. C. Snider (1915), who proposed the name Mayes for the basal formation of the Chester series in Mayes County, Oklahoma. The formation consists of the beds overlying the Boone unconformably and underlying the black shale of the Fayetteville formation. These beds, according to Snider's interpretation, were definitely the age equivalent of the Moorefield shale and Batesville sandstone of Arkansas. The Moorefield has since been definitely correlated as Meramecian by Mackenzie Gordon, Jr. (1944).

The Meramec age of the lower part of Snider's Mayes formation was recognized by George S. Buchanan (1927), who applied the name to the black, argillaceous Mississippian lime section of the subsurface, found in the lower part of the Caney shale. He extended the correlation south and west of the outcrop area to the Arbuckle Mountain region and north to the Kansas line. Buchanan apparently based his correlations, first, on the similarity of the lithology of the subsurface Mississippian rocks and the Mayes formation of Snider and, second, on the correlation of Aurin, Clark, and Trager (1921), which indicated that the black lime of the subsurface was younger in age than the Boone and that the Boone was removed by erosion and was missing west of the outcrop area.

TABLE 1

| | Mississippi Valley | Northeastern Oklahoma | | South-central Oklahoma | Ouachitas |
|---|---|---|---|---|---|
| CHESTER | | Pitkin<br>Fayetteville | | | Basal Stanley? |
| MERAMEC | Ste. Genevieve<br>St. Louis<br>Salem (Spergen)<br>Warsaw | Batesville<br>Ruddell<br>Moorefield | Mayes of Snider | Caney | Absent |
| OSAGE | Keokuk<br>Burlington<br>Fern Glen | Boone<br>St. Joe | "Mayes" of subsurface | Sycamore<br>Welden | Absent |
| KINDERHOOK | Chattanooga | Chattanooga | | Woodford | Upper Arkansas novaculite |

A few years later Ira H. Cram (1930) questioned the validity of Buchanan's correlation and offered some convincing arguments supporting his contention that the "Mayes" of the subsurface probably represented a remarkable change in facies of the Osage rocks from the white cherty limestones of the typical Keokuk-Burlington beds.

### CORRELATION AND THICKNESS OF MISSISSIPPIAN SUBSURFACE STRATA IN NORTHWESTERN ANADARKO BASIN

The four major subdivisions of the Mississippian are represented in the

# STATE OF OKLAHOMA

## AREAL DISTRIBUTION OF DIFFERENT OSAGE FACIES

ERWIN L. SELK
1947

Anadarko Basin and here attain their maximum thickness for this area.

Beds of Kinderhook age—the Woodford shale of this province—thin westward and pinch out entirely in the extreme northwestern part of the state, where the overlying Osage rocks rest upon Ordovician. The Woodford shale ranges in thickness from 0 to more than 500 feet.

Beds of Osage age have the lithologic characteristics of the Fern Glen formation in the lower part, and several facies of undifferentiated Burlington-Keokuk strata in the upper part. Their thickness varies from about 300 to 770 feet, except in areas of pre-Pennsylvanian truncation and along the Wichita Mountain front, where thicknesses in excess of 1,000 feet have been penetrated.

It is this series of beds which in Oklahoma has been correlated in various ways with the major subdivisions of the Mississippian and so engendered the existing confusion and controversy. The different facies of these strata, one of which is the "Mayes," will be discussed later in this paper.

The lithologic sequence of the beds in the Meramec series from the basal Warsaw formation up through the Ste. Genevieve formation is remarkably like the equivalents of the classic section in the subsurface of southwestern Illinois. This parallelism is so striking that the terminology of the Meramec formations in the classic section can be applied to these subsurface strata with assurance. The maximum thickness of these beds so far revealed in the subsurface is found in Beaver County, Oklahoma, where they are approximately 800 feet thick. These beds thin to the east. Because the section decreases from 600 feet to less than 100 feet across Dewey County, a pronounced pre-Chester unconformity is indicated.

The Chester overlies the Meramec, and its lithology is similar to that shown in the Illinois Basin section, with one disappointing exception—the well-defined sandstones of the type section have not been noted. Persistent fine-grained sandstones, however, occur in the lower part and are present higher in the section in those areas where the thickest Chester is found. Some of the limestone beds are similar to those of the Pitkin of the northeastern outcrops, but the black shales associated with the Pitkin occur only where the Chester changes facies along the north side of the Wichita Mountain front. Here it is approximately 2,400 feet thick, which is probably exaggerated because of steeply dipping beds, and overlies the Caney shales of Mississippian age.

The Chester age of the strata has been definitely established in southwestern Kansas from a study of the fauna obtained from numerous cores from exploratory wells. This determination was made by M. K. Elias[2] and as yet has not been published.

### FACIES AND LITHOLOGY OF THE OSAGE

There are four distinctive facies (fig. 1) of the Osage encountered in the subsurface of Oklahoma which in this paper are referred to as (1) "White Lime" facies, (2) "Siltstone" facies, (3) "Siliceous" facies, and (4) "Mayes-Sycamore" facies. These facies are most marked in the Keokuk-Burlington portion of the Osage series. The Fern Glen formation, where present, commonly consists of characteristic gray-green shale and buff, dense limestone, in descending order, and varies in thickness from a few feet to as much as 100 feet.

The "White Lime" facies consists of white to light-gray cherty limestone, commonly coarsely crystalline and cri-

[2] Personal correspondence.

designated the "Cowley formation." The age of the Cowley in Lee's interpretation is Meramecian, older than Warsaw. The abrupt change in lithology from the characteristic white limestone of the Keokuk-Burlington to the dark, silty, siliceous beds occupying the lower part of the Mississippian section throughout the southern tier of counties in Kansas was, according to Lee, a depositional fill, onlapping the margin of a dissected Osage terrain. Other reasons for his interpretation were based on a study of the insoluble residues and the persistence of a glauconite zone at the base of the Cowley, which thinned to the south toward the Oklahoma line in a direction away from the apparent source. The writer has noted that this glauconitic zone can be traced continuously from localities of the Cowley formation into the basal part of the "Mayes," where it is persistent.

## CONCLUSIONS

It is apparent that the "Siliceous" facies, the "Siltstone" facies, and the "Mayes-Sycamore" facies are stratigraphically equivalent to the Cowley formation.

The following facts must be considered in determining the age relationship between the white Osage limes and cherts and the Cowley formation: (1) both are overlain by characteristic Warsaw, which in outcrops is recognized as the basal formation of the Meramec; (2) there is no marked variation in relative thickness; and (3) the abrupt change in lithology is not unique.

A parallel situation exists in the subsurface strata of the Illinois Basin. The Keokuk-Burlington rocks grade abruptly into a gray calcareous siltstone and shale facies from northwest to southeast in the vicinity of the Du Quoin flexure. In the southern part of the Illinois Basin the character of these rocks is identical with the dark, cherty, siliceous limestones found in the "Siliceous" facies of northern Oklahoma and southern Kansas. These rocks are classified as Osagian throughout the Eastern Interior region.

The foregoing evidence surely justifies the contention that the "Mayes" of the subsurface in Oklahoma is Osagian.

ACKNOWLEDGMENTS.—The writer is indebted to the Stanolind Oil and Gas Company for permission to publish this report and for the use of subsurface data contained in its files and is also indebted to the various oil companies for permission to use the logs on their respective exploratory wells in the cross sections incorporated with this paper.

The writer is especially indebted to Robert H. Dott, of the Oklahoma Geological Survey, and John G. Bartrum, of the Stanolind Oil and Gas Company, for helpful suggestions in the preparation and editing of this report.

The writer wishes to express appreciation for the helpful discussions with various geologists of this area on matters relating to the problem, especially to M. K. Elias, of the University of Nebraska, Conservation and Survey Division, for permission to refer to his faunal determinations pertaining to the Upper Mississippian.

## REFERENCES CITED

AURIN, F. L.; CLARK, G. C.; and TRAGER, E. A. (1921) Notes on the subsurface pre-Pennsylvanian stratigraphy of the northern Mid-Continent oil fields: Am. Assoc. Petroleum Geologists Bull. 5, pp. 117–153.

BUCHANAN, G. S. (1927) The distribution and correlation of the Mississippian of Oklahoma: Am. Assoc. Petroleum Geologists Bull. 11, pp. 1307–1320.

CRAM, I. H. (1930) Oil and gas in Oklahoma: Oklahoma Geol. Survey Bull. 40qq, pp. 26–43.

GORDON, MACKENZIE, JR. (1944) Moorefield formation and Ruddell shale, Batesville district, Arkansas: Am. Assoc. Petroleum Geologists Bull. 11, pp. 1626–1634.

LEE, WALLACE (1940) Subsurface Mississippian rocks of Kansas: Kansas Geol. Survey Bull. 33, pp. 66–78.

SNIDER, L. C. (1915) Geology of northeastern Oklahoma: Oklahoma Geol. Survey Bull. 24, pp. 21–43.

# THE POSSIBILITY OF A LAND BRIDGE ACROSS NEBRASKA IN MISSISSIPPIAN TIME[1]

E. C. REED

Nebraska Geological Survey

ABSTRACT

The lithologic and paleontologic differences between the Mississippian sediments of the Mid-Continent region and those of the northern Rocky Mountain and Black Hills regions suggest either that the Mississippian seas were not directly continuous between these two regions or that sea connections were greatly restricted. The purpose of this paper is to present and analyze the available subsurface data in Nebraska and surrounding states in the light of this problem. The study is primarily lithologic.

Mississippian sediments are known to be widely distributed in the subsurface of much of Iowa and Kansas, in southeastern and extreme southwestern Nebraska, and in southeastern Colorado. These sediments are lithologically similar to those of the Mid-Continent outcrop areas. Likewise, Mississippian sediments are known to occur widely in the subsurface of much of Wyoming, in western and northwestern South Dakota, in northwestern Colorado, and in extreme northwestern Nebraska. These sediments are lithologically similar to those of the northern Rocky Mountain outcrop areas. However, Mississippian rocks seem to be absent in the subsurface in large areas between these two regions, and the Mississippian sediments of these two regions are lithologically dissimilar.

Therefore, it appears that there was either no direct sea connection between the Mid-Continent and northern Rocky Mountain regions during the Mississippian or that the sea connection was greatly restricted. Mississippian sediments could have been deposited between these two regions and removed by post-Mississippian erosion. It is unlikely, however, that great thicknesses were removed because of the apparent absence of good evidence suggesting facial changes. However, the pre-Pennsylvanian rocks are known to be deeply buried within much of the critical area, and the subsurface has not been thoroughly tested by drilling.

Mississippian sediments seem to be absent in the subsurface of a large part of Nebraska. This condition suggests the possibility that a land bridge extended across Nebraska during Mississippian time and that there was no direct sea connection between the Mid-Continent and the northern Rocky Mountain regions during the Mississippian. It is apparent (1) that many factors should be considered in arriving at a definite conclusion; (2) that the Mississippian cannot be treated as a single unit but that special considerations must be given to the possibility that sea connections may have existed at certain times during the Mississippian and not at other times; (3) that some areas have not been explored sufficiently for positive conclusions; and (4) that the subsurface and surface evidence in states adjoining Nebraska must be considered as important parts of the picture.

[1] Manuscript received December 26, 1947.

Although a number of deep wells have been drilled in Nebraska, many of those in critical areas have not been drilled deeply enough to reach Mississippian or pre-Mississippian rocks. Figure 1 shows the locations of test wells drilled to rocks of Mississippian and pre-Mississippian age. Note that large untested areas exist in northeastern, north-central, western, and southwestern Nebraska. However, there is a large area in the central part of the state where the Mississippian is undoubtedly absent, and the southward thinning of the Mississippian in northwestern Nebraska suggests that these rocks may be expected to be absent in much of the "panhandle" part of western Nebraska.

Whether or not the Mississippian is represented in the east-central Nebraska Basin is open to some question because of differences of opinion among qualified subsurface geologists. A succession of dense, very finely crystalline limestones

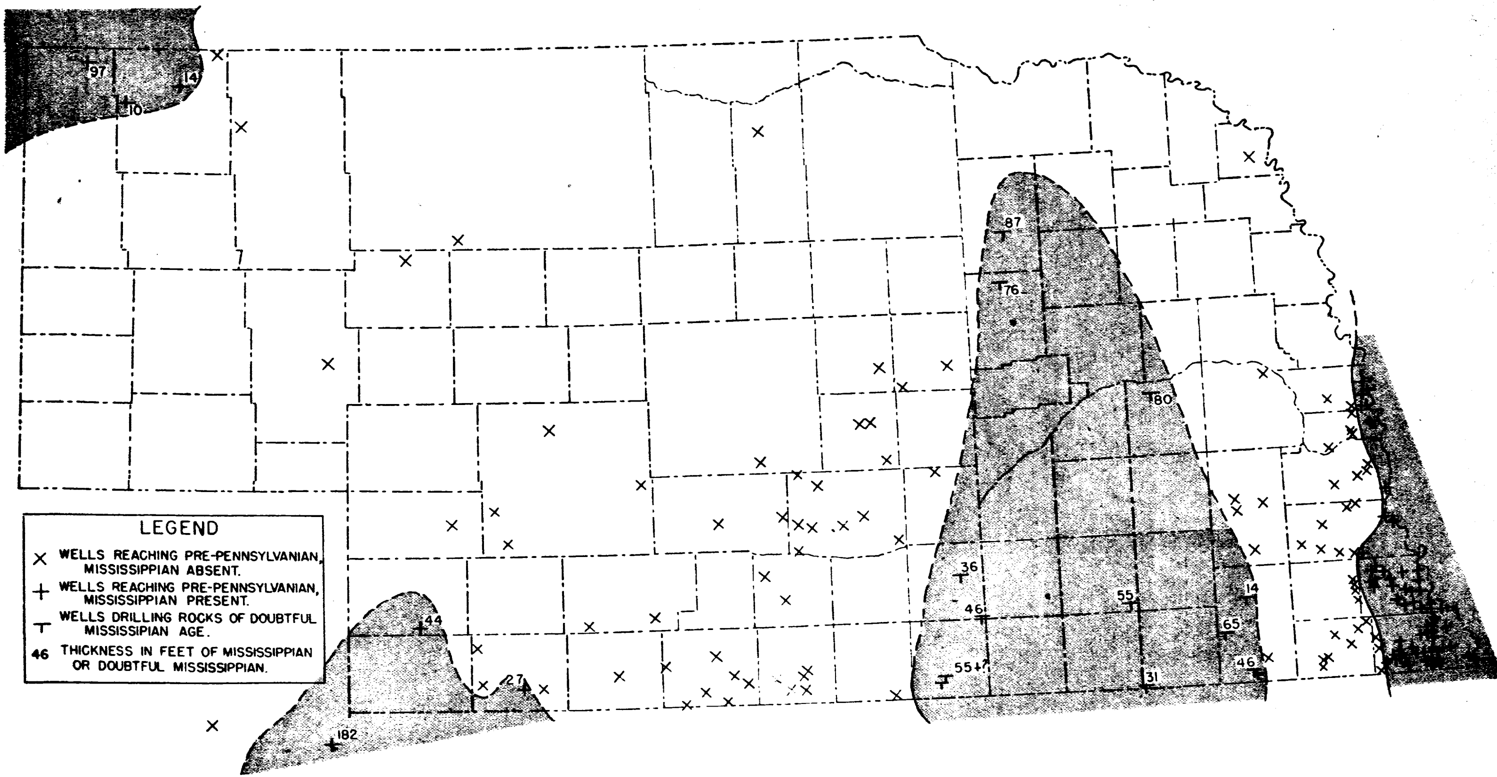



FIG. 1.—Map showing location of wells drilled in Nebraska to Mississippian or pre-Mississippian rocks. Areas shaded, where Mississippian is known to be present. Thickness of Mississippian penetrated shown, except in areas of concentrated drilling. Doubtful Mississippian rocks believed to be post-Cedar Valley Devonian by some subsurface geologists.

immediately underlies the Pennsylvanian in this area and rests, with apparent unconformity, upon relatively thin Devonian dolomites, probably of Cedar Valley age. These limestones are believed by some geologists to be Devonian, though post–Cedar Valley; but the writer believes that they may be of Kinderhook age and, therefore, should not be omitted from consideration. Their relation to overlying and underlying rocks is shown in figure 2. Rocks of undoubted Mississippian age have been penetrated in a number of wells in the Forest City Basin region in southeastern Nebraska, in several wells near the Kansas-Nebraska line in southwestern Nebraska, and in a few wells in the northwestern part of the state (fig. 1).

The dominant anticlinal structural features of Nebraska are the Cambridge Arch, which is a north-northwest extension of the Central Kansas Uplift and continues toward or into the Black Hills of South Dakota, and the Table Rock–Nehawka Arch, a north-south structural high, which is a northward continuation of the so-called "Nemaha Ridge" of Kansas. Post-Mississippian, pre-Pennsylvanian uplift took place along both these structural highs, and the Mississippian thins by truncation as the structurally high areas are approached. The thinning of the Mississippian by truncation toward the Table Rock–Nehawka Arch is shown in figures 3 and 4. In areas of close control the shale of Kinderhook age definitely thins to the northwest and north, and the individual formations in the Osage-Meramec show some north-northwest thinning. The beds of Chouteau age, however, preserve their rather uniform thickness where present and actually thicken in the eastern Nebraska Basin if the pre-Pennsylvanian, post–Cedar Valley rocks in that area should prove to be of Chouteau age.

If we eliminate the complications introduced by the Cambridge and Table Rock–Nemaha arches and assume that the Mississippian sea covered large areas of these uplifts and that the Mississippian rocks were removed by erosion prior to the deposition of Pennsylvanian sediments, there remains a broad northeast-southwest area devoid of Mississippian rocks which cannot be easily explained by post-Mississippian, pre-Pennsylvanian erosion along known lines of post-Mississippian uplift.

In comparing the relationships of the rocks in Nebraska with those in adjoining states (fig. 5), we note that this broad area lacking Mississippian rocks appears to continue northeastward into southeastern South Dakota, northwestern Iowa, and southern Minnesota and that Mississippian rocks appear to be absent in a broad area in east-central Colorado and southeastern Wyoming. This strongly suggests a land barrier between the Mid-Continent and the northern Rocky Mountain regions during much of Mississippian time. The sparse subsurface control in much of the critical area of western Nebraska and eastern Colorado precludes definite conclusions, however, as future deep drilling may encounter buried Mississippian outliers in structurally low areas.

It seems logical to conclude that it is very unlikely that great thicknesses of Mississippian rocks were deposited across Nebraska between the Mid-Continent and the northern Rocky Mountain regions and later removed. First, it is reasonable to expect that some of the "chance" drillings in this region would have encountered remnants of Mississippian rocks. Second, if large amounts of Mississippian rocks had been removed in early Pennsylvanian time, we should expect to find detrital materials from Mississippian sources in early Penn-

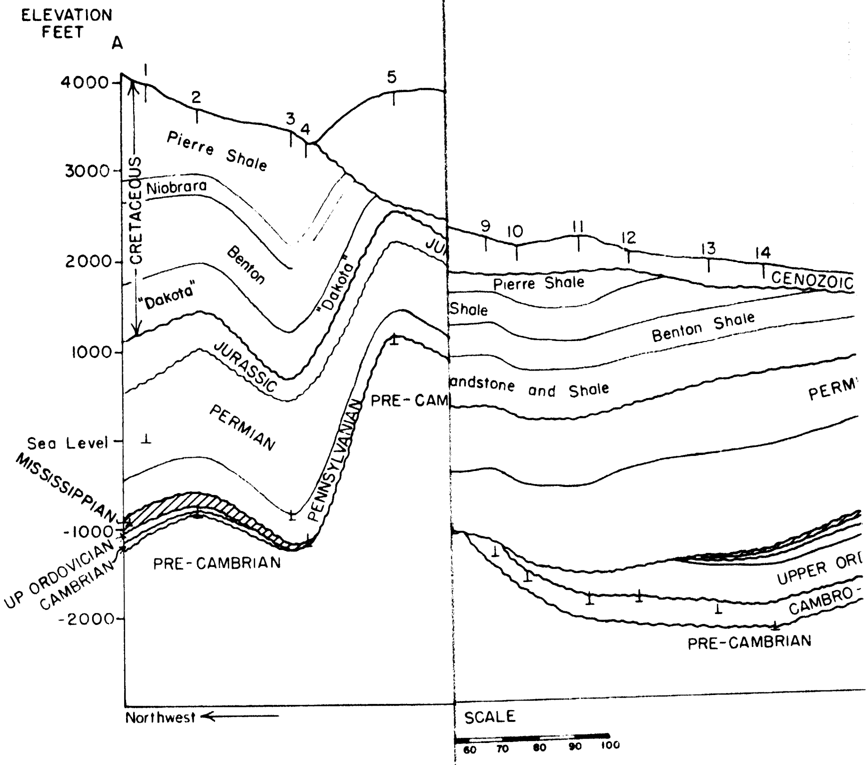
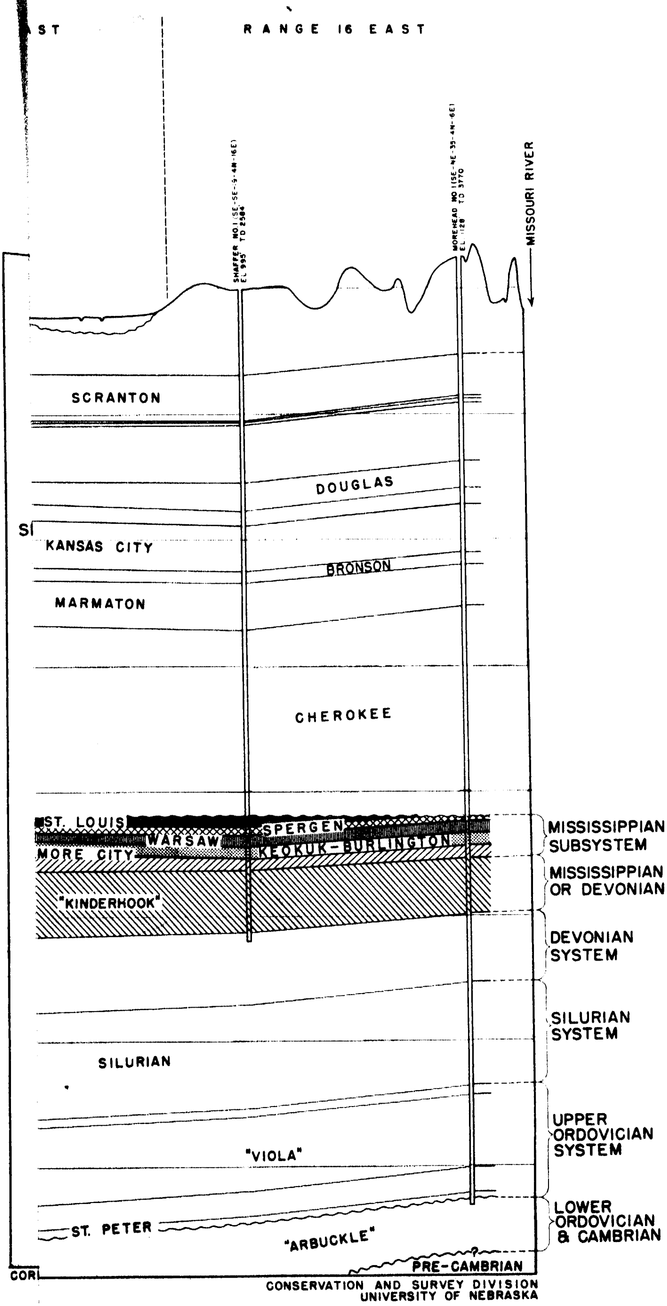



FIG. 2.—Northwest-southeast pr|rata (*cross-hatched areas*). Note principal structu Nehawka Arch (wells 18 and 19), ea|stern Nebraska Basin (wells 1–4).




ain and relation to Table Rock–Nehawka arch to north-
w

sylvanian deposits of the basinward parts of the region. However, re-worked Mississippian materials seem to be missing except along the more southerly extensions of the Cambridge and Table Rock–Nehawka arches.

In much of this area in Nebraska, Pennsylvanian rocks rest upon the pre-Cambrian. The considerable depth of weathering on the pre-Cambrian surface (more than 60 feet in some places) suggests a considerable time interval, probably longer than might be expected if the pre-Cambrian had been protected by Mississippian and older Paleozoic rocks and exposed to weathering for only shorter intervals of geologic time.

The post-Kinderhook Mississippian rocks of the northern Mid-Continent area seem to preserve their lithologic identity over wide areas in northern Kansas and southern Nebraska but tend to disappear northward by thinning and overlap. Moreover, the Mississippian of the northern Rocky Mountain region bears little or no lithologic similarity to the post-Kinderhookian Mississippian of the Mid-Continent region, nor does it show any tendency toward facial change in a southeast direction. Therefore, in view of all the evidence, it seems improbable that there was a post-Kinderhookian–Mississippian sea connection across Nebraska.

The evidence that a land barrier existed in this region throughout Kinderhookian time is not conclusive, however, because we cannot detect any marked northward or northwestward thinning of the Kinderhook limestones, except by trunca-

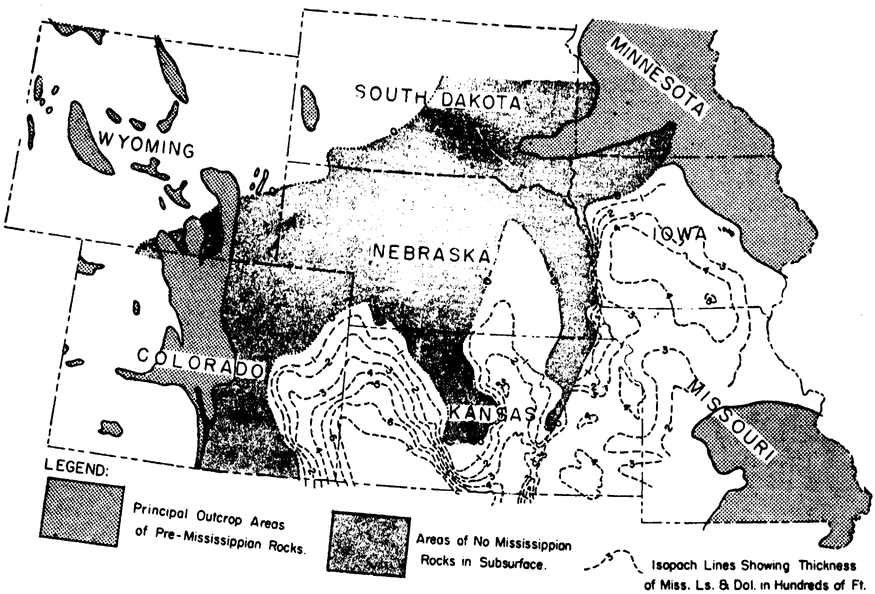



FIG. 5.—Generalized Mississippian distribution map of Nebraska and adjoining states based on published and unpublished data showing generalized subsurface thickness of Mississippian limestones and dolomites in eastern Colorado, Kansas, southern Nebraska, Missouri, and Iowa. Area of no Mississippian rocks generalized and includes small thickness of Mississippian in a few places near the area margins.

tion. Moreoever, if the doubtful limestone interval in the east-central Nebraska Basin (fig. 1) is truly Kinderhookian and not Devonian in age, some facial change between the Mid-Continent and the northern Rocky Mountain regions is indicated, as these rocks, especially in their northern extension, are not dissimilar lithologically to the Mississippian limestones of northwestern Nebraska and adjoining southwestern South Dakota.

ACKNOWLEDGMENTS.—The author has had the opportunity of checking all the subsurface evidence in Nebraska but has had to rely on published data in adjoining states, augmented by personal communications with survey personnel in Iowa, Kansas, South Dakota, and Wyoming. He is especially indebted to Dr. H. G. Hershey, state geologist of Iowa; Mr. Wallace Lee, of the United States Geological Survey and Kansas Geological Survey; Dr. C. L. Baker, of the South Dakota Geological Survey; and Dr. H. D. Thomas, state geologist of Wyoming, for assistance in their respective areas. He realizes that much pertinent subsurface information exists in the files of oil companies that have been engaged in regional studies, but no attempt has been made to obtain these data because much of it is confidential. He is also indebted to the authors listed in the appended bibliography for their important contributions.

## REFERENCES

BAKER, C. L. (1947) Deep borings of western South Dakota: South Dakota Geol. Survey Rept. of Inv. 57.

BALLARD, NORVAL (1942) Regional geology of the Dakota Basin: Am. Assoc. Petroleum Geologists Bull. 26, pp. 1557–1584.

BARB, C. F. (1946) Selected well logs of Colorado: Colorado School of Mines Quart., vol. 41, no. 1.

BRAINERD, A. E.; BALDWIN, H. L., JR.; and KEYTE, I. A. (1933) Pre-Pennsylvanian stratigraphy of the Front Range in Colorado: Am. Assoc. Petroleum Geologists Bull. 17, pp. 375–396.

——— and JOHNSON, J. H. (1934) Mississippian of Colorado: Am. Assoc. Petroleum Geologists Bull. 18, pp. 531–542.

CLAIR, J. R. (1943) The oil and gas resources of Cass and Jackson counties, Missouri: Missouri Geol. Survey and Water Resources, vol. 27, 2d ser.

CONDRA, G. E., and REED, E. C. (1943) Geological section of Nebraska: Nebraska Geol. Survey Bull. 14, 2d ser.

EDSON, F. C. (1945) Subsurface geologic cross section from Ford County to Wallace County, Kansas: Kansas Geol. Survey, Oil and Gas Inv., Preliminary Cross Section 1.

GREENE, F. C. (1945) Recent drilling in northwestern Missouri: Missouri Geol. Survey and Water Resources, Rept. of Inv. 1.

IOWA GEOL. SURVEY (1937) Geologic map of Iowa. Scale, 1:500,000.

LEE, WALLACE (1940) Subsurface Mississippian rocks of Kansas: Kansas Geol. Survey Bull. 33.

———; GROHSKOPF, J. G.; GREENE, F. C.; HERSHEY, H. G.; HARRIS, S. E., JR.; and REED, E. C. (1946) Structural development of the Forest City Basin of Missouri, Kansas, Iowa, and Nebraska: U.S. Geol. Survey Oil and Gas Inv., Preliminary Map 48, 7 sheets.

LOVERING, T. S., and JOHNSON, J. H. (1933) Meaning of unconformities in stratigraphy of central Colorado: Am. Assoc. Petroleum Geologists Bull. 17, pp. 353–374.

MCQUEEN, H. S., and GREENE, F. C. (1938) The geology of northwestern Missouri: Missouri Geol. Survey and Water Resources, vol. 25, 2d ser.

MAHER, J. C. (1946) Correlation of Paleozoic rocks across Las Animas Arch in Baca, Las Animas and Otero counties, Colorado: Am. Assoc. Petroleum Geologists Bull. 30, pp. 1756–1763.

——— (1946) Subsurface geologic cross section from Ness County, Kansas, to Lincoln County, Colorado: Kansas Geol. Survey Oil and Gas Inv., Preliminary Cross Section 2.

——— (1947) Subsurface geologic cross section from Scott County, Kansas, to Otero County, Colorado: Kansas Geol. Survey Oil and Gas Inv., Preliminary Cross Section 4.

MISSOURI GEOL. SURVEY AND WATER RESOURCES (1939) Geologic map of Missouri (revised). Scale, 1:500,000.

REED, E. C. (1946) The Boice shale, a new Mississippian subsurface formation: Am. Assoc. Petroleum Geologists Bull. 30, pp. 348–349.

U.S. GEOL. SURVEY (1925) Geologic map of Wyoming. Scale, 1:500,000.

——— (1932) Geologic map of the United States. Scale, 1:2,500,000.

VOLK, R. H. (1937) Present status of development in eastern Colorado: Oil and Gas Jour., vol. 36, no. 27, pp. 21–22, 32.

# SOME PROBLEMS OF MISSISSIPPIAN STRATIGRAPHY IN SOUTHWESTERN UNITED STATES[1]

ALEXANDER STOYANOW
University of Arizona

ABSTRACT

The southwestern Mississippian problems discussed include the relation of Lower Mississippian strata to their subjacent rocks, especially the problems involved in the evaluation of the Ouray limestone and the Percha formation as possible transitional units from the Devonian. Further problems are indicated in the relation of the Escabrosa limestone, with an impoverished brachiopod fauna, to the Redwall limestone, which is rich in brachiopods, and in the interrelation of the fossiliferous Lower Mississippian strata of Nevada, Utah, and Colorado. Although insufficient knowledge of the regional stratigraphic paleontology is a serious handicap for interpretation, the relation between the autochthonous forms of the southwestern Lower Mississippian and the species characteristic of the Mississippi Valley Basin suggest a westward migration from the latter region. However, the nature of the ways of communication is still problematical. For instance, the abundance of typical Mississippi Valley species and the lack of southwestern brachiopods in the Lower Mississippian of New Mexico do not favor a direct connection with the Redwall and the Escabrosa basins.

The Upper Mississippian Paradise formation of southeastern Arizona, most readily comparable paleontologically and stratigraphically to the standard Upper Mississippian of the Mississippi Valley, still has no other counterparts in the Southwest and is not easily traceable eastward through New Mexico and Texas or through Chihuahua and Coahuila in Mexico. The southernmost outpost of the comparable western unit, the Brazer limestone, with stratigraphically diffused paleontological indices and without *Archimedes* facies, is a long distance to the north, which suggests research problems in the intermediate areas, whereas the recent finds of Mississippian faunas in Sonora suggest the possibility of a southwestern orientation of a late Mississippian passageway in agreement with the pre-Mississippian paleogeographical setting and in common with the general direction of other Paleozoic outlets of that region.

## RELATION OF THE ESCABROSA AND REDWALL LIMESTONES IN ARIZONA AND OF THE LOWER MISSISSIPPIAN STRATA IN COLORADO AND OTHER SOUTHWESTERN STATES TO THEIR SUBJACENT ROCKS

The generally accepted standard stratigraphic units of the southwestern Lower Mississippian are the Escabrosa limestone in southeastern Arizona, the Redwall limestone in north-central Arizona, the Lake Valley formation in New Mexico, the Monte Cristo limestone in Nevada, the Leadville limestone in Colorado, and the Madison limestone in Colorado, Utah, and southeastern Idaho.

### ESCABROSA

In the southeastern part of Arizona the protean Devonian Martin limestone of Ransome (1904) is composed of several distinct stratigraphic units which may be readily correlated on paleontological evidence with the well-known standard Upper Devonian divisions in Iowa. Thus, in the Mescal Mountains south of Globe the strata between the Cambrian Troy quartzite and the Escabrosa limestone contain the fauna of the Independence shale of Iowa (with *Macgeea parva*, *Petrocrania famelica*, *Schizophoria amanaensis*, *Hypothyridina emmonsi*, *Spirifer strigosus*, etc.); at Superior, beneath the Escabrosa, is the Martin limestone proper, which is the Arizona equivalent of the Hackberry shale (Lime Creek) of Iowa as originally defined (with *Spirifer hungerfordi* and *Pachyphyllum woodmani*), whereas below the Martin limestone and separated from it by a cross-bedded sandstone are limestone beds with the index fossils of the Cedar Valley limestone of the Iowa sequence (*Sp. iowensis*, *Sp. cedarensis*, *Sp. euruteines*,

[1] Manuscript received February 2, 1948.

etc. [Stoyanow, 1936, p. 489]). There are also earlier fossiliferous Devonian strata between the Cedar Valley equivalent and the Troy quartzite below; but whether these beds contain the same assemblage of Independence fossils as is present in the Mescal Mountains has not been ascertained, and it is still unknown whether all three units of the Iowa Upper Devonian are represented in a single section anywhere in southeastern Arizona.

Stainbrook (1947, pp. 297–302) reports fossils of Percha age collected near the top of Mount Martin at Bisbee. He correlates the Percha formation of New Mexico with the Ouray limestone of southwestern Colorado (as redefined by Kirk, 1931, p. 222) but considers the Percha to be of early Mississippian age. Whatever the age of the Percha fauna—whether it is Devonian, Mississippian, or, like the Malevka-Murajevna, a transitional fauna—it is most probably not a time equivalent of the *Paurorhyncha endlichi* fauna of the Ouray limestone. *Paurorhyncha cooperi*, which Stainbrook cites both from the Percha formation of New Mexico and from the section of Mount Martin at Bisbee and which he compares with *P. endlichi*, has been located recently (Stainbrook, 1947, pl. 47, fig. 1, holotype only) by W. B. Loring, 95 feet below the base of the Escabrosa limestone of the Swisshelm Mountains in the extreme southeastern part of Arizona, near the New Mexico border. However, the lowest known occurrence of *P. endlichi*, near Burch on Pinal Creek, northwest of Globe,[2] is only a few inches above the topmost reef of *Hexagonaria* (*Acervularia davidsoni* of various authors). Because both the reef and *P. endlichi* occur in the same lithological unit, a Mississippian age for this species is not possible.

[2] The collected specimen is identical with the type illustrated by White (1883, pl. 33, fig. 4*a* only).

The late Devonian strata in southeastern Arizona show marked variability, a rapid change in faunal succession, and a spotty distribution. With the exception of the Superior area, the equivalents of the Cedar Valley and of Hackberry are rarely present in the same section. The Devonian strata and the Escabrosa are apparently conformable. The apparent conformity may be an illusion arising from the striking lithological and physiographic contrast between the easily eroded, slope-forming Devonian strata and the bold relief of the cliff-forming Escabrosa limestone. Certainly, however, there was no communication through Colorado between Arizona and the Iowa Devonian basins. The patchy distribution of the Independence fauna in the Southwest is further illustrated by its unexpected discovery in isolated areas, such as New Mexico near Alamogordo (Stainbrook, 1935) and the Mescal Mountains of Arizona. On the other hand, a communication between the Devonian seas of Arizona and southwestern Colorado seems to have been established at the close of Devonian time (as inferred from the distribution of the *P. endlichi* fauna), and it is certainly pertinent to inquire, therefore, whether there actually are transitional faunas and gradational deposition between the Devonian and the Mississippian within such units as the Ouray limestone and the Percha formation.

### REDWALL

The Redwall limestone—the Lower Mississippian stratigraphic unit of north-central Arizona—rests on the uneven surface of different older strata. Observations indicate that the basin in which the

Redwall formed continued to the northwest and that the thickness of the Redwall appreciably decreases both to the northeast and to the southeast. Wooddell (1927) was the first to notice the unconformable contact between the Devonian Jerome formation and the Mississippian Redwall limestone near Jerome, Arizona. The regional unconformity is more apparent if broader areas are considered. In the vicinity of Jerome certain younger Devonian strata, like the Island Mesa beds (Stoyanow, 1936, p. 500; 1942, p. 1271), are present only locally. The overlap of the Redwall limestone on the Cambrian platform is well shown in the Grand Canyon sections.

### OURAY, PERCHA, AND LEADVILLE

In central Colorado the Lower Mississippian Leadville limestone is readily distinguished from the Devonian strata. Kirk (1931) has established a Devonian-Mississippian unconformity in the triangle outlined by Glenwood Springs, Salida, and Leadville. He introduced the term "Leadville" for the Lower Mississippian strata of that area, extended it also to cover the Lower Mississippian part of the Ouray limestone, and restricted the latter name to the Devonian part.[3] However, in the San Juan region and especially in the type area of the Ouray limestone the unconformity has not been so clearly demonstrated. An uninterrupted continuity of deposition without any appreciable change in lithology from the Devonian to the Mississippian has been described in a number of publications (Cross and Howe, 1905, 1907; Cross and Ransome, 1905; Cross, 1910). It is interesting to note in this connection that, according to the original description of the type Ouray limestone by Spencer (1900, p. 126), the thickness of strata between the Devonian part of the limestone and the upper part, which was believed to belong in the Upper Carboniferous, amounts to only a few feet. This led Spencer to the conclusion that either there is a gap in the sequence which represents Mississippian time or else some unfossiliferous strata in the lower part of the limestone, but above the fossiliferous Devonian zone, may be of Mississippian age. In other words, an idea of disconformity was introduced with the original description. Certain lithological differences between the Devonian and the Mississippian parts of the sequence at Ouray were pointed out by Kindle (1909, pp. 6–7, 13), who was unable to find Devonian fossils at the type locality. According to Kirk (1931, p. 233), the type area ". . . is the place where it is most difficult to separate the two stratigraphic units. The exposures are either in cliffs difficult of access or badly overgrown with timber." The fossiliferous localities in the vicinity are few, and in some areas no fossils have been found. In an earlier paper Burbank (1930, pp. 158–161) stated that the Ouray limestone at Ouray is divisible into two lithological units, of Devonian and Mississippian age, respectively. The Devonian part is approximately 68 feet thick. The only fossiliferous Devonian zone located is about 15–20 feet above the base and, according to Dr. Kirk, contains a characteristic Devonian Ouray fauna. The upper limit of the Devonian was placed at the base of a thin-bedded, blue-gray limestone with nodules of chert. In some

[3] Kirk thus reverted to the original definition of Spencer (1900, p. 126): "The formation name, Ouray limestone, is proposed for the only member of the section which is definitely shown by its fossils to be of Devonian age, from the prominent occurrence in the vicinity of Ouray at the junction of the Canyon creek with the Uncompaghre River."

places a limestone breccia occurs in this unit, in others the base of the Mississippian seems to be marked by a sandstone.

It seems that the break between the Upper Devonian and the Lower Mississippian in Colorado was smallest in the San Juan region, and it is probable that in certain parts of southwestern Colorado deposition continued without interruption. From the standpoint of stratigraphical paleontology a locality northeast of Animas, Colorado, is of special interest. Here Endlich (1876, pp. 211–214, Station 48 on the map, p. 412) collected the types of *P. endlichi*, forms identical with the plesiotypes associated with the Upper Devonian faunas in Arizona and probably different from the majority of specimens referred to that species in the literature.

As noted above, Stainbrook (1947, p. 302) believes the Percha fauna to be most closely related to the Devonian fauna of the Ouray limestone. Observing that Kindle (1909) showed a similarity between the Ouray and Percha faunas, he says: "If the Percha be Mississippian, the Ouray is Mississippian also, and the Devonian is but scantily represented in Colorado." I think, however, that this correlation of the two faunas has not been satisfactorily proved. The fossils listed from the Ouray of Colorado and the close relation of the beds with typical *P. endlichi* to strata with the characteristic Hackberry shale faunas in certain parts of Arizona oppose this view. The abundance of productids in the Percha fauna described by Stainbrook, and especially such genera as *Avonia*, *Buxtonia*, *Krotovia*, and *Echinoconchus*, altogether unknown from the Ouray of Colorado, in which Productidae are represented only by productellids (Johnson, 1945, p. 41), strongly favors a younger age for the Percha if it is considered a single time unit. It is significant, however, that nearly all productids described by Stainbrook came from the Bella member, i.e., the top of the Percha formation, and the possibility that there are faunas of two ages in the Percha cannot be excluded.[4] On the other hand, if the Percha fauna is accepted as a unit (biochron) of relatively short range, then Percha time may be included in the hiatus between the Ouray and the Leadville in central Colorado. The presence of a similar break at Ouray, where neither fossils nor interruption in sequence is definitely known between the *P. endlichi* and *Spirifer centronatus* zones, is more questionable. Stainbrook's views bring back, on a broader basis, the old controversy which constitutes one of the current Mississippian problems of the Southwest.

### OTHER INTERCORDILLERAN STATES

In southeastern Nevada the Lower Mississippian Monte Cristo limestone rests on the Devonian Sultan limestone without a perceptible unconformity (Hewett, 1931, pp. 15–17). Farther northeast, in the Muddy Mountains, the base of the Lower Mississippian is indicated by an irregular surface of the limestone assigned to the Devonian (Longwell, 1928, p. 28), whereas in central and eastern Nevada and western Utah the Lower Mississippian strata rest on different formations essentially of middle and late Devonian ages (Westgate and Knopf, 1932, pp. 17, 19–20; Nolan, 1943, pp. 153–154). In the Gold Hill mining district, Utah, the base of the Madison is obscured by a thrust fault (Nolan, 1935, pp. 21, 24–27, figs. 3–5),

[4] Stainbrook (1947, p. 298) himself states that his interest in the described material was essentially paleontological and that "while collecting from the Percha no stratigraphic studies were made other than to locate the top and bottom of the formation, if exposed, and to verify the position of the beds with respect to adjacent formations."

but an unconformity is inferred on stratigraphic evidence. In the Stockton-Fairfield quadrangles there is a definite erosional break between the Madison limestone and the supposed Jefferson (Devonian) dolomite (Gilluly, 1932, p. 22). In southeastern Idaho the base of the Madison limestone is not exposed (Mansfield, 1927, p. 60).

### STRATIGRAPHIC EVALUATION OF THE SOUTHWESTERN LOWER MISSISSIPPIAN FAUNA

Progress in paleontological stratigraphy of the southwestern Mississippian has been greatly handicapped by the lack of adequate studies and analyses of the fossil material used for correlations. The onerous and ever increasing work of identifying the fossils collected by the field parties of the United States Geological Survey was for years executed by the late G. H. Girty, without opportunity for sufficient description and illustration, and therefore resulted in deductions often not wholly acceptable to others. Correlation of strata from the Mexican border to Idaho by a single paleontologist, of no matter how great experience, could not be infallible and has resulted in wide generalizations and probably some errors. The practice of borrowing paleontological illustrations from publications on strata in certain cases many hundreds of miles distant and of doubtful relation to the formations of a given locality, the fossils of which they were supposed to characterize, is actually harmful to detailed stratigraphic analysis. Such conditions could not contribute much to the progress of paleontological stratigraphy and paleogeography of a vast region, and the need of research to bring these branches of regional geology to the level of other geological studies is considerable.

#### ESCABROSA LIMESTONE

Table 1 shows species of correlative value from the Escabrosa limestone which also occur in the lists from sections studied in the Southwest and in southeastern Idaho.

TABLE 1*

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| *Syringothyris "carteri"* | ...... | Sp.? | ...... | Aff. | ...... | × | ...... |
| *Spirifer centronatus* | × | × | ? | × | × | × | × |
| *Leptaena analoga* | ...... | × | ...... | × | ...... | ...... | × |
| *Chonetes loganensis* | × | × | × | × | × | × | × |
| *Rhipidomella thiemi* | ...... | × | ...... | ...... | ...... | ? | ...... |
| *Linoproductus ovatus* | × | × | × | × | × | ...... | × |
| *Camarotoechia metallica* | × | × | ...... | × | Cf. | × | × |
| *Triplophyllum?* several sp. | × | × | × | × | ...... | ? | × |
| *Euomphalus luxus* | ...... | ...... | ? | × | × | ...... | × |

* Index fossils of the Escabrosa limestone in relation to the faunal lists from other characteristic Lower Mississippian sections of the Southwest and southeastern Idaho. 1 = Pioche district, Nevada; 2 = Good Springs, Nevada; 3 = Gold Hill, Utah; 4 = Stockton-Fairfield, Utah; 5 = Wasatch and Uinta, Utah; 6 = Colorado; 7 = southeastern Idaho.

The presence of *Syringothyris* in the Escabrosa limestone (Ransome, 1904, pp. 42–54) has not been corroborated by subsequent workers. It is not improbable that "*Syringothyris carteri*" in Ransome's list of fossils from Bisbee belongs to *Syringospira*, a genus established by Kindle (1909, pp. 28–30, pl. 7, figs. 8–8*d;* pl. 8, figs. 1–1*a*) on material from the Percha formation at Hillsboro, New Mexico, three years after the publication

of Ransome's monograph on the Bisbee quadrangle and therefore probably not known to Girty at that time. *Syringospira prima* Kindle has been cited by Stainbrook (1947, p. 298) from the strata near the top of Mount Martin at Bisbee, that is, close to the base of the Escabrosa limestone at its type locality. As seen from table 1, *S. typa* (*S. carteri* auctorum) has been definitely indicated only in the Crested Butte quadrangle, Colorado (Johnson, 1945, table 7).

Among the varieties of *Sp. centronatus*, a form shorter and wider and also less mucronate than any other of the described types referred to Winchell's species (White, 1877, pl. 5, figs. 8*a*–8*b*; Hall and Whitfield, 1877, pl. 4, figs. 5–6), deserves special consideration because of its definite stratigraphic position and wide distribution. In the Escabrosa limestone this variety invariably occurs with abundant "small cup corals" which were referred by Girty to *Triplophyllum* and *Menophyllum* in the material collected from both the Escabrosa and the Redwall limestones (Ransome, 1904, p. 50; 1916, pp. 143, 147, 152; Darton, 1925, p. 65). These genera, together or separately, are listed from the Mississippian formations of Nevada (Longwell, 1928, pp. 30–31; Hewett, 1935, pp. 19–20), Utah (Gilluly, 1932, pp. 23–24; Nolan, 1935, pp. 26–27), Colorado (Girty, 1903, p. 271; Johnson, 1945, table 7), and also the Madison limestone of southeastern Idaho (Mansfield, 1927, p. 60). Easton (1944) has shown that *Triplophyllum* proper is of Onondaga age, whereas the Lower Mississippian *Triplophyllites* has a long range, through the Tournaisian and Viséan in Europe. The stratigraphic paleontology of these little corals of the Madison sea, therefore, is another problem of the southwestern Mississippian.

The forms so abundant in the Escabrosa and present in the beds with *Sp. centronatus* belong neither in *Triplophyllum* nor in *Triplophyllites*. They are closely related to certain species of the Mississippi Valley (Easton, 1944) but in many cases show a reduction and even total absence of transverse endothecal structures. Whether or not the indicated morphological difference observed in the Arizona forms also involves a different stratigraphic position (Chouteau-Burlington in the Mississippi Valley) may be questioned; but the Burlington age of the upper part of the Escabrosa limestone is supported by the presence of a blastoid, *Orophocrinus stelliformis* (Stoyanow, 1936, p. 507), at present known from widely separated localities (Santa Catalina and Veekol Mountains, a distance of over 100 miles). It is significant, also, that at the base of the Paradise formation in the Chiricahua Mountains, in the extreme southeastern part of Arizona, Chouteau-Burlington brachiopods, like *Sp. gregeri*, occur below the *Sp. centronatus* zone (Hernon, 1935, p. 658).

The Lower Mississippian "*Leptaena analoga*" is generally interpreted as a species with a larger shell than the earlier "*L. rhomboidalis*." The Escabrosa form is small (10 mm. long and 18 mm. wide). It is closer to *L. convexa* Weller and more resembles the pre-Mississippian species.

One of the outstanding characteristics of the Escabrosa is the paucity of productids. If "*Productus semireticulatus*" of Ransome's (1904, p. 50) collection from Bisbee is the Pennsylvanian *Dictyoclostus americanus* (which is quite probable because the Pennsylvanian Naco limestone rests with perfect apparent conformity on the Escabrosa limestone), that leaves only 2 species—*Linoproductus ovatus* and *D. mesialis* (Stoyanow, 1936, p. 505). In marked contrast the Redwall limestone contains 15 species, and the Lower Mis-

sissippian formations of neighboring states have 7 or 8 species, including related forms. There are 7 species in the Monte Cristo limestone of Nevada; 7 in the Stockton and Fairfield quadrangles in Utah; 7 in Colorado; and at least 8 in the Madison limestone of southeastern Idaho. On the other hand, only 1 species has been reported from the Madison limestone in the Gold Hill mining district of Utah, and 4 specifically unidentifiable forms have been listed from the Muddy Mountains of Nevada. The Lower Mississippian of New Mexico has 7 species.

Separation of the Escabrosa into zonal units may be done successfully, as it has been for the Avon series in England, on the basis of vertical and widespread areal distribution (from the Mexico border to Lake Roosevelt) of the rugose corals (especially several species of *Clinophyllum*, *Hadrophyllum*, *Dipterophyllum*, *Homalophyllites*, and *Cyathaxonia*) profusely abundant in many parts of the Escabrosa.

### REDWALL LIMESTONE

The Redwall has not been adequately explored from the standpoint of paleontological stratigraphy in the Grand Canyon area, except for casual collecting on the trails cutting through inaccessible cliffs and sheer walls of the massive limestone. What is known of the Redwall fauna is based on collections gathered along the southwestern rim of the Colorado Plateau. C. E. Wooddell (1927) made a fine collection of fossils in the Jerome area (see Stoyanow, 1936, p. 512), which is now on deposit at the University of Arizona; and, more recently, Gutschick (1943) collected in the same area. Wooddell's collection contains 175 carefully studied and described species, mostly Kinderhook to Burlington in age, with Keokuk age less certainly established. The difference in the number of species compared to that of the Escabrosa is impressive. The Redwall has 10 species of brachiopods in common with the Escabrosa, 32 with the Lower Mississippian of the Mississippi Valley, 9 with the Caballero formation of New Mexico (Laudon and Bowsher, 1941), 15 with the Alamogordo member of the Lake Valley formation, and 6 each with with Argente and Dona Ana members of the same formation. On the other hand, the Escabrosa has only 7 brachiopod species in common with the Lower Mississippian of the Mississippi Valley Basin, 2 with the Caballero formation, none with the Alamogordo and Argente members of the Lake Valley formation, and 3 which occur in the Dona Ana member.

In the Lower Mississippian of the other states under discussion, the Monte Cristo limestone of Nevada contains (barring related forms) 10 brachiopod species of the Lower Mississippian of the Mississippi Valley; the Madison limestone in the Stockton and Fairfield quadrangles, Utah, has 8 such species, and in the Gold Hill mining district of the same state only 1 definitely identified species; the Madison limestone in southeastern Idaho contains 7 species of the Mississippi Valley.

Assuming equal ability of collectors and a similar preservation and frequency of fossils in different parts of the "Madison sea," it is easy to see that the depositional basin of the Redwall limestone was an exceptionally favorable habitat for the brachiopods. The latter are selected in this discussion because they are more easily identified and so cause fewer errors in stratigraphic interpretations than do other less adequately studied Lower Mississippian invertebrates of the Southwest. There are no known assemblages of well-preserved and easily determinable

crinoids, similar to those in the Dona Ana member of the Lake Valley formation of New Mexico, in other areas. The corals, as a group, have not yet been adequately studied.

Whereas many Lower Mississippian brachiopods of the Mississippi Valley occur in the strata of the southwestern sea, certain index species of the southwestern Lower Mississippian are unknown in the Mississippi Valley. This, most probably, suggests a westward migration. However, it is difficult to indicate the paths of migration and channels of communication. The Lake Valley formation of New Mexico has over fifty brachiopod species and the Redwall limestone has over thirty species which have been described from the Mississippi Valley. The other formations in the neighboring states have much less. Yet a direct communication between Redwall and Lake Valley basins does not appear plausible on present evidence. Neither the Lower Mississippian fauna of Colorado, which contains eleven brachiopod species in common with the faunas of the Mississippi Valley and New Mexico, nor the Escabrosa fauna, with seven such species, nor the geographically intermediate area of Clifton-Morenci in Arizona near the New Mexico border, with only one such species, indicates an unimpeded interchange between the faunas of the two basins. Significant also is the absence in the Lake Valley formation of such a characteristic southwestern brachiopod as *Sp. centronatus*.[5] As in the cases of the Upper Devonian and the Upper Mississippian, a communication must be sought elsewhere.

[5] The form from Michigan and Iowa described under this name by Weller (1901, p. 163, pl. 14, figs. 3–4; 1914, p. 323) is not conspecific with *Sp. centronatus* Winchell of the Southwest, which apparently was autochthonous and does not occur east of Colorado and Arizona.

## EVALUATION OF THE UPPER MISSISSIPPIAN AREAS IN THE SOUTHWEST

### PARADISE FORMATION, BRAZER LIMESTONE, AND WOODMAN FORMATION

The Paradise formation of Arizona (Stoyanow, 1926, p. 316; 1936, p. 508; 1942, p. 1273, pl. 5, fig. *h;* Hernon, 1935, p. 653) is of importance for interpretation of the Upper Mississippian stratigraphy of the Southwest because of its place in the sequence and because its relation to the type regions is much clearer than that of other comparable formations farther north. Over sixty species in the Paradise fauna, not counting varieties or specimens of doubtful affinity, are strictly identical with those described and illustrated by Stuart Weller (1914) and others from the Mississippi Valley and, in lesser measure, from Arkansas and Oklahoma (Moorefield-Pitkin). The clarity of interpretation increases upward from the equivalents of the St. Louis to the Chester. The areal extent of the Paradise formation is very limited; especially limited is the Chester *Archimedes* facies, the bryozoan fauna of which has been recently described by Condra and Elias (1944). It is interesting that the late Dr. Girty considered the *Archimedes* and *Pentremites* faunas, wherever abundant, to be regional types, restricted chiefly to the eastern half of the continent (Mansfield, 1927, p. 69).

Among other stratigraphic units of the Southwest with firmly established Upper Mississippian faunas, two are of special interest, namely, the Woodman formation in the Gold Hill mining district, Utah (Nolan, 1935, p. 27), and the Brazer limestone of Utah and southeastern Idaho (Mansfield, 1927, p. 63). The latter is particularly important because it includes faunas related to all facies from the Salem to the Chester in the single regional sequence.

In the Woodman formation, Nolan distinguished strata of an eastern facies, which rest unconformably on the Devonian, and strata of a western facies, which directly succeed the Madison limestone. Unfortunately, nearly all the brachiopods listed from the Woodman formation (with the exception of five species) are only related forms. Girty believed that the faunas of the Woodman formation and of the overlying Ochre Mountain limestone are not sharply distinguishable and contain assemblages too small for exact interpretation. The Upper Mississippian strata of the district are correlated with western stratigraphic units, but no comparison with the standard sections of the Mississippi Valley is available. In the Stockton-Fairfield quadrangles farther northeast (Gilluly, 1932, p. 25), the nature and relation of the collected fossils are very similar, except that *Linoproductus brazerianus* is reported from the same formation (Humbug limestone) with a form related to *Sp. centronatus*, and neither species has been found in the Upper Mississippian beds of the Gold Hill district.

Next to the Paradise, the Brazer limestone of southeastern Idaho is the most important formation for the evaluation of the southwestern Upper Mississippian. Despite Girty's statement (Mansfield, 1927, p. 63) that most of the collections taken from the Brazer ". . . appear no more related to one group than to another, or, by lacking individual character, show no close relation to any," many authentic species are illustrated in Mansfield's monograph and make a sound interpretation possible. Girty's major conclusions are: (1) the Chester fauna, though present mainly at the top of the Brazer, is not typical Chesterian because, for example, the *Archimedes* facies is absent; (2) some facies are related to stratigraphic occurrences, others are not; (3) some productids and spiriferids "are distinctly alien to the faunas of the typical Chester, indeed, they represent types not found in the Mississippi Valley above the Keokuk"; and (4) there are many forms in the part of the Brazer correlated with the Spergen which do not occur in the Spergen.

Other valuable observations by Dr. Girty need not be enumerated. Correlations with the Spergen and the Chester were made on either identical or related species. It is important to note that, because the faunas of the Brazer are considerably diffused vertically and are rather patchy laterally, collateral studies of beds farther east and west must be made before the true nature of these faunas is properly understood. There are many problems; for example, Girty regarded the so-called "alien" species of the Brazer, "*Productus*" *brazerianus*, *Martinia lata*, etc., as suggestive of older species in the Lower Mississippian of Asia, Continental Europe, and England. It seems that such forms may equally well be considered to be related to certain Pennsylvanian species of the Ural Mountains. Girty's *M. lata* bears a close resemblance to *M. semiglobosa* Tschern. (Tschernyschew, 1902, pl. 17, figs. 6–9 and 12–13), including the observed and discussed similarity to the Reticulariinae. "*P.*" *brazerianus* Girty (in Butler *et al.*, 1920, p. 643, pl. 53, figs. 1–2*a*), with its small beak, also seems rather to resemble the less inflated but wide-hinged forms of the Uralian "*P. cora*" (Tschernyschew, 1902, pl. 35, fig. 1; pl. 54, figs. 1–5) rather than *Gigantella gigantea* (Martin), which has unusually large and broad umbonal region (Davidson, 1857, pls. 39 and 40). The relation of the Brazer species to "*P. cora*" and "*P.*" *giganteus* was discussed at length by Girty himself. It is

quite probable that this species is autochthonous. *Spirifer haydenianus* Girty (in Mansfield, 1927, p. 416, pl. 24, figs. 18–21), which Girty compared to *S. striatus* (Martin) of the European Mississippian (Davidson, 1857, pl. 2, figs. 12–21; pl. 3, figs. 2–6) also does not seem to be related to the latter species any more than to "*Sp. striatus*" from the Pennsylvanian of the Urals (Tschernyschew, 1902, pl. 40, figs. 5*a–c*).

## PALEOGEOGRAPHIC RELATIONS AND PROBLEMS

Granting that the widespread dispersal of a marine fauna takes considerably less time than any evolutionary changes that may be incurred during the migration, a satisfactory correlation of strata containing faunas with identical or closely related species in different, and even very remote, regions is justified. However, the identity of remote faunas requires an explanation of their paleogeographic relationship. The paleogeographer may freely pencil in seas and continents, but, as the geographic distance decreases, his responsibility for the correctness of his interpretations increases in geometrical progression.

Any regional paleogeographic interpretation for a given unit of time must take account of factors inherited from the previous periods and the regional changes of previous times. Southwest and south of the Colorado Plateau in Arizona the Paleozoic strata exposed in a great many outcrops invariably consist of a uniform sequence of limestone beds, with the exception of certain areas in the vicinity of the ancient pre-Cambrian upland in central Arizona. The thin conglomerate at the base of the Pennsylvanian, present everywhere except in extreme southeastern Arizona, and the weak, often gypsiferous and shaly, members of the Permian (Stoyanow, 1942), long remained unnoticed, probably because of their insignificant volume in the dominating mass of Paleozoic limestones. The remarkable uniformity and the apparently perfect conformity of the strata of five different Paleozoic periods (the Middle and Upper Ordovician, Silurian, Lower and Middle Devonian, beds are absent in all sections; the Lower Ordovician and Upper Mississippian strata are present only in limited areas near the New Mexico border) have not been obscured either by the major time intervals or by the post-Paleozoic orogenic movements. The widespread Upper Cambrian–Upper Devonian contact is especially impressive.

The relative scarcity of clastic materials in the southeastern half of the state, the absence of marginal belts of clastic rocks, and the overwhelming predominance of limestone in the Paleozoic strata all suggest that no major Paleozoic geosynclines have ever crossed Arizona and that the ancient schistose-granitic platform, "Mazatzal Land" of the writer (Stoyanow, 1936, 1942), remained essentially undisturbed and an area of comparatively low relief throughout the Paleozoic.

Nolan (1928, p. 154; 1935, p. 24; 1943, p. 153) has pointed out the principal characteristics of the Mississippian Cordilleran geosyncline in Nevada and Utah, which in early Mississippian time had a northeasterly trend approximately parallel to a line from the Death Valley region in California to Gold Hill in Utah. The geosyncline later spread farther westward. It is reasonable to suppose that the Redwall basin connected to the northwest with the geosyncline approximately in the region of strongly folded mountains in southwestern Utah.

It is more difficult to demonstrate

communication channels for the Escabrosa and the Paradise seas. There is little in common in the faunal lists of the Escabrosa and the Lake Valley of New Mexico, and there is slight probability of a direct connection between the two basins. There remain the alternatives of northwestern and southeastern communications. It is possible that there were impeded and brief connecting channels between the Escabrosa and the Redwall basins either directly across the barrier of the Mazatzal quartzite in central Arizona or through an unknown area now under the younger beds of the plateau. The impoverished brachiopod fauna of the Escabrosa may have resulted from such conditions. An alternate or concurrent channel may have passed through Mexico. There is very strong evidence of a Paleozoic trough and a westward portal in northern Sonora (Stoyanow, 1942, pp. 1263, 1272, 1279). I have some Lower Mississippian fossils collected by my Mexican colleagues from the limestone beds near Bizani in the Altar Valley which suggest an Osage age, and recently Cooper and Arellano (1946, p. 606) reported the presence of Lower and Middle Mississippian in the same area. However, judgment on the relation of the Bizani fauna to the Escabrosa (compare Stoyanow, 1942, p. 1272) would be premature.

Concerning the possibility of a southern outlet of the Cordilleran geosyncline of late Mississippian time in the Great Basin, Nolan (1943, p. 153) states: "The later Mississippian sea extended considerably farther westward into Nevada, to the Inyo Mountains, California. Whether it was joined with the sea of the same age in central and northern California by a passage to the north or to the south is not yet known." I have called attention in two articles to the absence of Paleozoic sediments in southwestern Arizona, where the only exposed bedrock is pre-Cambrian schist, granite, and subrecent volcanics, and have also pointed out the known localities where Paleozoic fossils have been redeposited in post-Paleozoic, often poorly consolidated, strata (Stoyanow, 1942, pp. 1272, 1278–1279). I referred to the extreme southwestern part of Arizona as "Altar headland of Mazatzal Land" on the assumption that the schistose-granitic basement extends into Sonora. In my opinion the geological evidence of the presence and stability of the southwestern headland is rather strong. It appears that in southwestern Arizona there is a virgation of the schistose and granitic ranges nearly around a center of resistance, in striking contrast with the monotonous northwestern trend of the ranges in southeastern and central Arizona. This virgation is well indicated in the Geologic Map of Arizona prepared by the Arizona Bureau of Mines in co-operation with the United States Geological Survey, and I have attempted to illustrate it diagrammatically (Stoyanow, 1942, p. 1257, fig. 1).

There are no undoubted Paleozoic deposits in west-central Arizona south of the Harquahala Mountains (Darton, 1925, pp. 215–223; Stoyanow, 1942, p. 1278), and all supposed evidence to the contrary deals with the finds of redeposited fossils on the periphery of the discussed area in southwestern Arizona. Recently Cooper and Arellano (1946, p. 608) questioned the Altar headland (they have had the advantage of doing field work in Sonora, where I have never been) and point out the presence of metamorphic rocks and granitic intrusions. They state that, although they are unable to determine the age of these rocks, their observations and published papers on the Paleozoic of Sonora give no support to the concept of such a

"headland." The occurrence of intrusive rocks in the Altar-Magdalena region of Sonora, especially where such rocks have intruded and mineralized limestone is fairly well known to the mining geologists of Arizona. Cooper and Arellano fail to cite any reference to the area between the Altar-Magdalena region and southwestern Arizona. Does the schistose-granitic platform stop abruptly at the international border, and, if not, how far and in what direction does it extend into Sonora? Information that I have from traveling geologists suggests that the westward extent is considerable; whether similar rocks extend southeastward neither Cooper and Arellano nor I know; and my theory, based on the difference of Middle Cambrian trilobite faunas found in southeastern Arizona and the Altar-Magdalena region, respectively, seems to be satisfactorily grounded. It should also be mentioned that at present the evidence in favor of the existence of Paleozoic seaways in southwestern Arizona is very insufficient. According to Russell Wheeler (1936), the Cambrian Bright Angel shale extends from the Grand Canyon area to the headwaters of the East Verde River, much farther south than I had supposed (Stoyanow, 1942, p. 1262, pl. 5, fig. *a*). A couple of years ago E. D. Wilson, geologist of the Arizona Bureau of Mines brought to Tucson a slab from the Harquahala Mountains in which I identified Cambrian trilobites. In the same area and in a redeposited condition 40 miles to the west, McKee (1947, p. 282) collected trilobites with *Glossopleura* (identified by Cooper) in the assemblage. Citing the finds of *Glossopleura* in Sonora (Stoyanow, 1942; Cooper and Arellano, 1946), McKee concluded: "Thus, it seems definite that in early Middle Cambrian time a seaway was continuous from western Mexico through Arizona into the Grand Canyon area." Inasmuch as McKee (1947, p. 288) believed that the lithological similarity between the Cambrian of Harquahala and the Grand Canyon indicated a close relation between these areas, it seems more probable that the Harquahala trilobites belong in the Cambrian Basin of the Grand Canyon area, there being no evidence of Cambrian strata on the schistose-granitic platform of southwestern Arizona. There have been no other attempts to postulate Paleozoic seas in that region of Arizona.

Bearing in mind the above discussion of the general paleogeographic setting in early Paleozoic time and also taking cognizance of the absence of the Ordovician and Silurian in Arizona (excepting the presence of the Beekmantown in the Clifton-Morenci quadrangle and in the Dos Cabezas Mountains along the New Mexico border) and northern Sonora, it is easier to interpret the nature of late Paleozoic rocks north of the international border and south of it in the Magdalena-Altar region of Sonora. All evidence of the Paradise fauna is lost within 60–80 miles west and northwest of the type locality. At Bisbee, in the Escabrosa Ridge and in the Naco Hills, an important break is observed between the Lower Mississippian and the Pennsylvanian. The same relationships, with or without a dividing conglomerate, are present in all border mountains to the west—the Huachuca, the Whetstone, the Empire, the Santa Rita, and the Patagonia Mountains. There was no depositional basin along the international border in Paradise time.

Hernon (1935, p. 659) thought that there might have been an eastern way of communication between the Paradise Basin and the eastern sources of the Upper Mississippian fauna across New

Mexico, but the more recent work of Laudon and Bowsher (1941) shows the absence of faunas younger than those of the lower Burlington formation of the Mississippi Valley in the Mississippian sections of that state. In the United States there are no known Upper Mississippian strata between the Paradise Basin and the area with the Helms formation in western Texas (Nelson, 1940, p. 165). Even the western Texas occurrence is in doubt because Condra and Elias (1944, p. 6) concluded, after examining the United States Geological Survey collection of *Archimedes* from the Helms area, that some of the specimens indicate Lower Pennsylvanian rather than Chester age.

Therefore, the finding of some part of the Paradise fauna and formation in the El Tigre area in Sonora and south of Douglas, Arizona, by Imlay (1939, pp. 1731–1732) is of particular interest. Cooper, who examined Imlay's collection, correlated the strata at El Tigre with the Ste. Genevieve formation of the Mississippi Valley and (probably) with the lower part of the Paradise formation. I know of no evidence that there was a direct eastward communication with the Mississippi Valley Basin south of the international border through Chihuahua and Coahuila. On the other hand, Cooper and Arellano (1946, p. 610) believe that they have located Upper Mississippian strata in the Magdalena-Altar region near Bizani; as is the case with the El Tigre area, the strata studied are correlated with the Ste. Genevieve of the Mississippi Valley and also with the lower part of the Paradise formation. This evidence, meager as it may seem, is the only indication of the regional distribution of the Paradise fauna in the Southwest. It is significant that the direction indicated by these finds *is to the west* and considerably southwest from the belts with the Mississippian strata in Arizona.

## REFERENCES CITED


BURBANK, W. S. (1930) Revision of geologic structure and stratigraphy in the Ouray district of Colorado, and its bearing on ore deposition: Colorado Sci. Soc. Proc., vol. 12, no. 6, pp. 151–232.

BUTLER, B. S.; LOUGHLIN, G. F.; HEIKES, V. C.; and others (1920) The ore deposits of Utah: U.S. Geol. Survey Prof. Paper 111.

CONDRA, G. E., and ELIAS, M. K. (1944) Study and revision of *Archimedes* (Hall): Geol. Soc. America Special Paper 53.

COOPER, G. A., and ARELLANO, A. R. V. (1946) Stratigraphy near Caborca, northwest Sonora, Mexico: Am. Assoc. Petroleum Geologists Bull. 30, pp. 606–611.

CROSS, C. W. (1910) Description of the Engineer Mountain quadrangle, Colorado: U.S. Geol. Survey Geol. Atlas, Engineer Mountain Folio 171.

——— and HOWE, ERNEST (1905) Description of the Needle Mountains quadrangle, Colorado: U.S. Geol. Survey Geol. Atlas, Needle Mountains Folio 131.

——— ——— (1907) Description of the Ouray quadrangle, Colorado: U.S. Geol. Survey Geol. Atlas, Ouray Folio 153.

——— and RANSOME, F. L. (1905) Description of the Rico quadrangle, Colorado: U.S. Geol. Survey Geol. Atlas, Rico Folio 130.

DARTON, N. H. (1925) A résumé of Arizona geology: Univ. Arizona, Arizona Bur. Mines Bull. 119.

DAVIDSON, THOMAS (1857) A monograph of British Carboniferous brachiopoda, vol. 2, pt. 5.

EASTON, W. H. (1944) Corals from the Chouteau and related formations of the Mississippi Valley region: Illinois State Geol. Survey Rept. Inv. 97.

ENDLICH, S. F. (1876) Report on the San Juan district, Colorado: U.S. Geol. and Geog. Survey Terr. 8th Ann. Rept. (Hayden), pp. 181–240.

GILLULY, JAMES (1932) Geology and ore deposits of the Stockton and Fairfield quadrangles, Utah: U.S. Geol. Survey Prof. Paper 173.

GIRTY, G. H. (1903) The Carboniferous formations and faunas of Colorado: U.S. Geol. Survey Prof. Paper 16.

GUTSCHICK, R. G. (1943) The Redwall limestone (Mississippian) of Yavapai County, Arizona: Plateau, vol. 16, no. 1, pp. 9–11.

HALL, JAMES, and WHITFIELD, R. P. (1877) Paleontology: U.S. Geol. Expl. 40th Par. Rept., vol. 4, pt. 2, pp. 197–302.

HERNON, R. M. (1935) The Paradise formation and its fauna: Jour. Paleontology, vol. 9, pp. 653–698, pls. 80–82.

HEWETT, D. F. (1931) Geology and ore deposits of the Goodspring quadrangle, Nevada: U.S. Geol. Survey Prof. Paper 162.

IMLAY, RALPH (1939) Paleogeographic studies in northeastern Sonora: Geol. Soc. America Bull. 50, pp. 1723–1744.

JOHNSON, H. J. (1945) A résumé of the Paleozoic stratigraphy of Colorado: Colorado School of Mines Quart., vol. 40, no. 3.

KINDLE, E. M. (1909) The Devonian fauna of the Ouray limestone: U.S. Geol. Survey Bull. 391.

KIRK, EDWIN (1931) The Devonian of Colorado: Am. Jour. Sci., 5th ser., vol. 22, pp. 222–240.

LAUDON, L. R., and BOWSHER, A. L. (1941) Mississippian formations of Sacramento Mountains, New Mexico: Am. Assoc. Petroleum Geologists Bull. 25, pp. 2107–2160.

LONGWELL, C. R. (1928) Geology of Muddy Mountains, Nevada, with a section through the Virgin Range to the Grand Wash Cliffs, Arizona: U.S. Geol. Survey Bull. 798.

MCKEE, E. D. (1947) Paleozoic seaways in western Arizona: Am. Assoc. Petroleum Geologists Bull. 31, pp. 282–292.

MANSFIELD, G. R. (1927) Geography, geology, and mineral resources of part of southeastern Idaho, with descriptions of Carboniferous and Triassic fossils by G. H. Girty: U.S. Geol. Survey Prof. Paper 152.

NELSON, L. A. (1940) Paleozoic stratigraphy of Franklin Mountains, west Texas: Am. Assoc. Petroleum Geologists Bull. 24, pp. 157–172.

NOLAN, T. B. (1935) The Gold Hill mining district, Utah: U.S. Geol. Survey Prof. Paper 177.

——— (1943) The Basin and Range Province in Utah, Nevada, and California: U.S. Geol. Survey Prof. Paper 197-D.

RANSOME, F. L. (1904) The geology and ore deposits of the Bisbee quadrangle, Arizona: U.S. Geol. Survey Prof. Paper 21.

——— (1916) Some Paleozoic sections in Arizona and their correlation: U.S. Geol. Survey Prof. Paper 98-K.

SPENCER, A. C. (1900) Devonian strata in Colorado: Am. Jour. Sci., 4th ser., vol. 9, pp. 125–133.

STAINBROOK, M. A. (1935) A Devonian fauna from the Sacramento Mountains near Alamogordo, New Mexico: Jour. Paleontology, vol. 9, pp. 709–714.

——— (1947) Brachiopoda of the Percha shale of New Mexico and Arizona: *ibid.*, vol. 21, pp. 297–328.

STOYANOW, A. A. (1926) Notes on recent stratigraphic work in Arizona: Am. Jour. Sci., 5th ser., vol. 12, pp. 311–324.

——— (1936) Correlation of Arizona Paleozoic formations: Geol. Soc. America Bull., vol. 47, pp. 459–540.

——— (1942) Paleozoic paleogeography of Arizona: *ibid.*, vol. 53, pp. 1255–1262.

——— (1948) Molluscan faunule from Devonian Island Mesa beds, Arizona (in press).

TSCHERNYSCHEW, T. (1902) Die oberkarbonischen Brachiopoden des Ural und des Timan: Com. géol. Mém. 16, no. 2.

WELLER, STUART (1914) The Mississippian Brachiopoda of the Mississippi Valley Basin: Illinois State Geol. Survey Mon. 1.

WESTGATE, L. G., and KNOPF, ADOLPH (1932) Geology and ore deposits of the Pioche district Nevada: U.S. Geol. Survey Prof. Paper 171.

WHEELER, R. B., and KERR, A. R. (1936) Preliminary report on the Tonto group of the Grand Canyon, Arizona: Grand Canyon Nat. History Assoc. Bull. 5, pp. 1–16.

WHITE, C. A. (1877) Report on the invertebrate fossils collected in portions of Nevada, Utah, Colorado, New Mexico, and Arizona: U.S. U.S. Geol. Survey W. 100th Mer. Rept., vol. 4, pt. 1.

——— (1883) Contributions to invertebrate paleontology 6: certain Carboniferous fossils from the western states and territories: U.S. Geol. and Geog. Survey Terr. 12th Ann. Rept., pt. 1, pp. 119–141.

WILLIAMS, S. J. (1943) Carboniferous formations of the Uinta and northern Wasatch Mountains, Utah: Geol. Soc. America Bull. 54, pp. 591–624.

WOODDELL, C. E. (1927) The Mississippian fauna of the Redwall limestone near Jerome, Arizona: Thesis submitted in partial fulfilment of the requirements for the degree of Master of Science in the Graduate College of the University of Arizona.

# MISSISSIPPIAN-PENNSYLVANIAN BOUNDARY PROBLEMS IN THE ROCKY MOUNTAIN REGION[1]

JAMES STEELE WILLIAMS
United States Geological Survey

ABSTRACT

A variety of paleontologic and stratigraphic problems are presented by rocks near the Mississippian-Pennsylvanian boundary in the central and northern Rocky Mountains. Stratigraphic sections of these rocks show diverse interpretations of fundamental concepts of stratigraphy and paleontology. In many places where Upper Mississippian rocks directly underlie Pennsylvanian rocks it is difficult to determine the precise location of the boundary between these units. Formations that straddle the boundary are very useful and satisfactory over large areas. Most geologists use various types of lithologic criteria to distinguish formations, but some appear to rely mainly on faunal data, unconformities, or attempts to trace prominent beds. More uniformity in criteria than now exists for the delimitation of formations is desirable. Surface and subsurface formations should conform to the same definition. Critical paleontologic studies of several common species and genera, if based on a large number of specimens, might help solve the boundary problem. More correlations based on several lines of paleontologic evidence and less reliance on a few index fossils would also help. Larger and more varied collections of well-preserved fossils stratigraphically located are needed from critical areas. Additional stratigraphic work in this region should be of a detailed nature and should preferably be done in connection with detailed mapping. Ecologic and paleogeographic factors merit more attention. The age significance of unconformities has perhaps been overestimated generally.

## INTRODUCTION

As in many other parts of the United States, the Mississippian and Pennsylvanian rocks of the Rocky Mountain region present many unsolved problems. These problems relate to all stratigraphic zones from the base of the Mississippian to the top of the Pennsylvanian. A group of problems that involve beds at or near the Mississippian-Pennsylvanian boundary are especially interesting because they not only show places at which the geological data are sadly deficient but also involve interpretations and differences in viewpoints on fundamental principles of paleontology and stratigraphy.

All students of Carboniferous problems, especially those who have themselves worked in the Rocky Mountains, will agree that much geologic work needs to be done there. The type of work most needed, in the writer's opinion, is not, however, reconnaissance work but detailed work, whereby the investigator becomes well informed on a single small problem or spends considerable time on a large problem. There is, however, also room for broadly interpretive work. Considering the vast area of the Rocky Mountains underlain by Carboniferous rocks and the difficulty of access of many of the exposures, a very creditable amount of knowledge of the stratigraphy and paleontology has existed for a long time; but not all of it is published, and much that is published is in papers concerned also with general and economic geology, with which papers many stratigraphers appear to be unfamiliar. This knowledge must be considered by anyone starting work in the Rocky Mountains.

Problems in the Rocky Mountain Carboniferous (not all of which will be solved or even reviewed in this paper!) range from the need for more and better fossils, carefully collected with respect to their geographic locations and stratigraphic horizons, to the need for reviews, and perhaps reappraisals, of some of the fundamental hypotheses and definitions used in stratigraphy and paleontology.

[1] Published by permission of the Director, U.S. Geological Survey. Manuscript received February 24, 1948.

Among these last-named are such things as definitions of various rock and time units and the applications of these definitions in the field; hypotheses of, and factors in, the correlation of strata; and theories of species definition in paleontology. Despite the two hundred and thirty or more years of the existence of the science of stratigraphy and stratigraphic paleontology, many disagreements exist in the application, if not in the definition, of many of the fundamental or near-fundamental concepts upon which the daily work of the stratigrapher and stratigraphic paleontologist is based.

Whether one considers the Mississippian-Pennsylvanian boundary a systemic, subsystemic, or series boundary depends on the definitions of a system, a subsystem, and a series to which one subscribes and on the applications (or interpretations) of these definitions in particular regions and with particular sequences of rocks; also involved are the uses or underlying purposes that one has in mind for each of the units, the general usage throughout the world, the degree of reliance and degree of fineness of intercontinental correlations of the particular units of rocks under consideration, and the breadth of experience one has with the rocks involved. All these are variable, and there is certainly adequate room for justified disagreements in the weights and interpretations given each of the above factors and for disagreement in the rank assigned to the units called "Mississippian" and "Pennsylvanian." A definite agreement is not necessary, and it would be outside the scope of this paper to present arguments for or against any specific conclusion. The writer considers that the Mississippian-Pennsylvanian boundary is an important boundary in the United States (more so in some regions than in others) and believes it to be an important time and time-rock boundary, as distinct from a lithologic boundary. It may (and does) happen to coincide with distinct lithologic changes in some places but not with important lithologic changes in others. It coincides with an unconformity in some regions and not with a recognizable unconformity in others. It is a practical boundary for mapping in some places, and in others it is not. Nevertheless, this boundary is one of the more important ones in the United States.

### POSITION OF MISSISSIPPIAN-PENNSYLVANIAN BOUNDARY

All who are familiar with the general geology of the central and northern Rocky Mountains know that, broadly speaking, the Mississippian rocks there constitute a sequence mainly of limestones, whereas the Pennsylvanian rocks constitute a dominantly sandstone or "quartzite" sequence. Between the dominantly limestone sequence of the Mississippian and the dominantly sandstone sequence of the Pennsylvanian there lies a series of thin and in many places alternating beds of sandstones, shales, thin limestones, cherts, and other kinds of rock. In many places this series of rocks contains red or purple beds, material from which stains associated beds and at many exposures the whole series has a reddish tinge. In many places the Mississippian-Pennsylvanian boundary is within this series of rocks, some of the beds being Mississippian and others Pennsylvanian. In other places, however, the Mississippian-Pennsylvanian boundary, as determined by fossils, appears to coincide with a lithologic boundary. The Mississippian-Pennsylvanian boundary is placed within a series of alternating thin-bedded rocks—a nonresistant series —not only in the area here discussed but

in a far wider area in the western part of the United States.

#### FORMATIONS INVOLVED

*Early practice.*—The variable beds between the Mississippian limestones and the Pennsylvanian sandstones or quartzites have, in the area under discussion, been placed in different formations in different parts of the area. In western and central Montana and in northwestern Wyoming they were generally assigned to the lower part of the Quadrant formation and widely, but not universally, considered Mississippian in age. In west-central and central-northern Wyoming and in parts of Montana contiguous to northern Wyoming, they were placed in the Amsden formation, which has from 1906 (Darton, 1906, p. 5), two years after the time of the proposal of the name "Amsden" for the beds, been generally considered to be of both Mississippian and Pennsylvanian age. In southeastern and eastern Idaho and contiguous parts of western Wyoming and Utah, the lower beds of the sequence were for a long time placed in the upper part of the Brazer limestone and the upper beds in the lower part of the Wells formation. In mapping begun in 1931 in the Afton quadrangle, southeastern Idaho and southern Wyoming, but as yet unpublished, W. W. Rubey and the writer grouped the beds together in a single mapping unit, to which a field name has been applied pending decision as to which of the available names to use. In north-central Utah the beds were put in the Morgan formation, which was considered by its namer, Eliot Blackwelder (1910, p. 530), to be Pennsylvanian in age, but which may contain Mississippian beds in its lower part. The writer has collected Pennsylvanian fossils from the type section and other exposures of the Morgan, and Mississippian fossils from beds that might be considered Lower Morgan. The name "Morgan" has been extended into the Cottonwood–American Fork region of the Wasatch Mountains and to other areas in this part of the Wasatch Mountains and has also been used at several places in the Uinta Mountains.

In central-western Utah, the nonresistant unit of alternating shales, limestones, and sandstones in which the Mississippian-Pennsylvanian boundary occurs has little red material. At this place the unit was called the Manning Canyon shale by Gilluly (1932, pp. 31–34). The name Manning Canyon has been used also for units of approximately the same age in eastern Nevada and at other places in central Utah, including at least one area in the Wasatch Mountains, near Provo (Baker, 1947).

The above paragraphs give the general usage as of about 1930 (fig. 1). This shows that in most places the Mississippian-Pennsylvanian boundary was frankly acknowledged not to be a practical mapping (i.e., formational) boundary but to lie within a formation. Exceptions to this were found, however, in the southeastern Idaho area, where attempts (abandoned by W. W. Rubey and the writer) were made to map the Mississippian-Pennsylvanian contact as the Brazer-Wells contact and, to a degree, in the area of the typical Morgan formation, where all of the Morgan was considered to be of Pennsylvanian age.

*Recent work.*—Much work has been done in the central and northern Rocky Mountains since 1930. The United States Geological Survey has had many field parties working in various parts of the region. Renewed interest in stratigraphic problems has been shown by some of the state geological organizations and especially by geologists on the faculties of col-

leges and universities in the area. Very important contributions have been made by the faculties and student bodies of the many summer camps maintained by midwestern and eastern universities in different parts of the Rocky Mountains.

some of the identifications are as yet provisional. Most of the sections measured by Geological Survey parties have been published. Unfortunately, wartime and other duties have prevented the writer from compiling and publishing many of

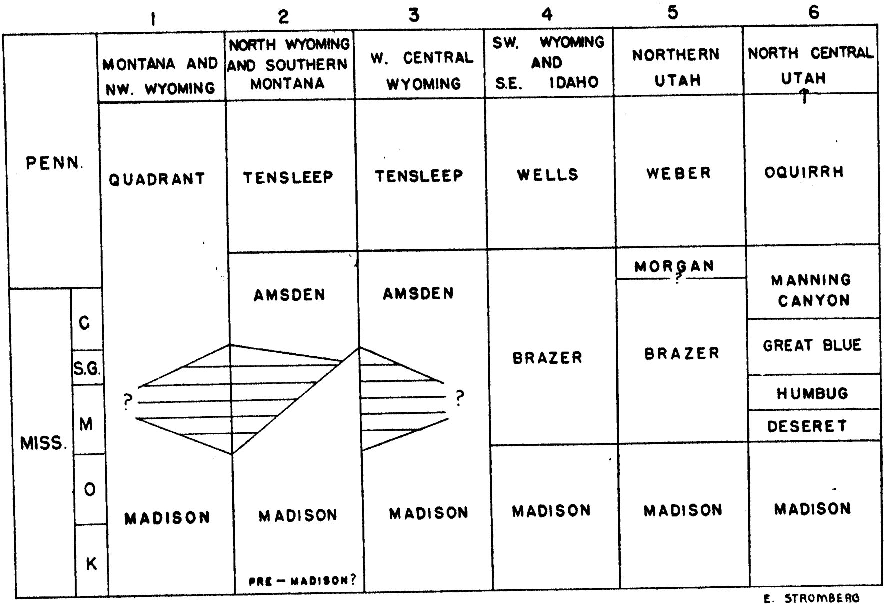



FIG. 1.—Correlation chart showing widespread usage as of about 1930. The stratigraphic usage shown for several of the areas on the chart is still the preferred and most satisfactory, but it has changed in other areas. The diagonal lines at the base and top of the cross-lined areas represent an attempt to show that both the top of the Madison limestone and the base of the Amsden formation probably are of different ages in different places, or are thought to be by different geologists. The abbreviations are for the well-known Mississippi Valley subdivisions of the Mississippian.

The writer has aided nearly every United States Geological Survey party that has worked in this area since 1930 with its problems in Carboniferous rocks and, in addition, while engaged in stratigraphic projects of his own, has measured many sections in areas not worked in detail by the Survey mapping parties. Fossils have been systematically, though not always adequately, collected from nearly all these sections, and most of these have been identified, although the sections that he has measured independently and from studying carefully all the collections of fossils in his hands.

SELECTED STRATIGRAPHIC SECTIONS

The total stratigraphic work done on the Mississippian-Pennsylvanian boundary problem by all the geologists who have worked in this region is so great and the number of stratigraphic sections is so large that only a small proportion of them can be discussed in this paper. Con-

sequently, a few sections have been selected to show the trends in each of several areas. It is hoped that these sections will reveal general tendencies in procedure that can be evaluated later and will show some of the specific deficiencies of knowledge, disagreements in philosophies, and other problems that exist in connection with the stratigraphy and paleontology of the Mississippian-Pennsylvanian boundary in the Rocky Mountains. The sections given here were selected because they are more or less typical of the areas or because they show fossil occurrences, lithologic features, or terminology that the writer believes are of interest. The writer has personally visited the area of each section cited from the various publications. Several of the sections have been examined in the field with the men who measured them.

### SECTION IN THE BIG SNOWY MOUNTAINS CENTRAL MONTANA

The section in the Big Snowy Mountains of central Montana (sec. 1, fig. 2) is condensed from one published by Scott in 1935 (p. 1024). It is given to indicate changes in usage that have gained wide acceptance in a part of the area where the nonresistant alternating beds were for a long time placed in the Quadrant formation. The locality is "on north flank of Big Snowy Mountains, sec. 6, T. 12 N., R. 20 E."

Reeves in 1931 (p. 140) recognized four units in the Quadrant formation which he described as follows: (1) an upper thin-bedded fossiliferous limestone interbedded with red shale, 100–200 feet thick; (2) a unit of red, brown, and black shales and cross-bedded sandstones, 300–400 feet thick; (3) a 500-foot unit of variegated calcareous shale, with a few thin limestones and including a predominantly green shale near the middle; (4) a basal unit of yellow sandstone, sandy shale, and gypsum, 150–200 feet thick. Girty, who examined the fossil collections made by Reeves, identified fossils from the middle and lower part of the upper limestone unit that, though not definitely diagnostic, were considered to be upper Mississippian in age (either Ste. Genevieve or Chester or both), but he considered a collection from near the top of this limestone unit at another place to be clearly Pottsville (Pennsylvanian). The writer has collections from this limestone, but he has not yet been able to study them or to re-examine Girty's collections.

The main sandstone or quartzite, to which the name "Quadrant" is now generally restricted and which is probably younger than any of the beds in Reeves's section, is represented only by a few outliers, if at all, in the Big Snowy area, but considerable thicknesses of this sandstone occur in other parts of Montana. In many places Jurassic beds rest on Reeves's upper thin-bedded limestone of the Quadrant formation. In 1935, Scott described some fifteen or more stratigraphic sections in Montana, including the one here cited in the Big Snowy Mountains. Like Reeves, he divided the Quadrant into four units in the Big Snowies, but his units did not coincide precisely with those of Reeves. For his two lower units he brought in the names Kibbey and Otter, names used by Weed for members of the Quadrant in the Little Belt Mountains. To a unit composed in large part of "black petroliferous shales and sandstones" immediately below the upper limestone unit of Reeves, he gave the name Heath formation. For the upper limestone unit he brought in Darton's (1904, p. 396) Amsden from central-northern Wyoming. Scott considered the Kibbey, Otter, and

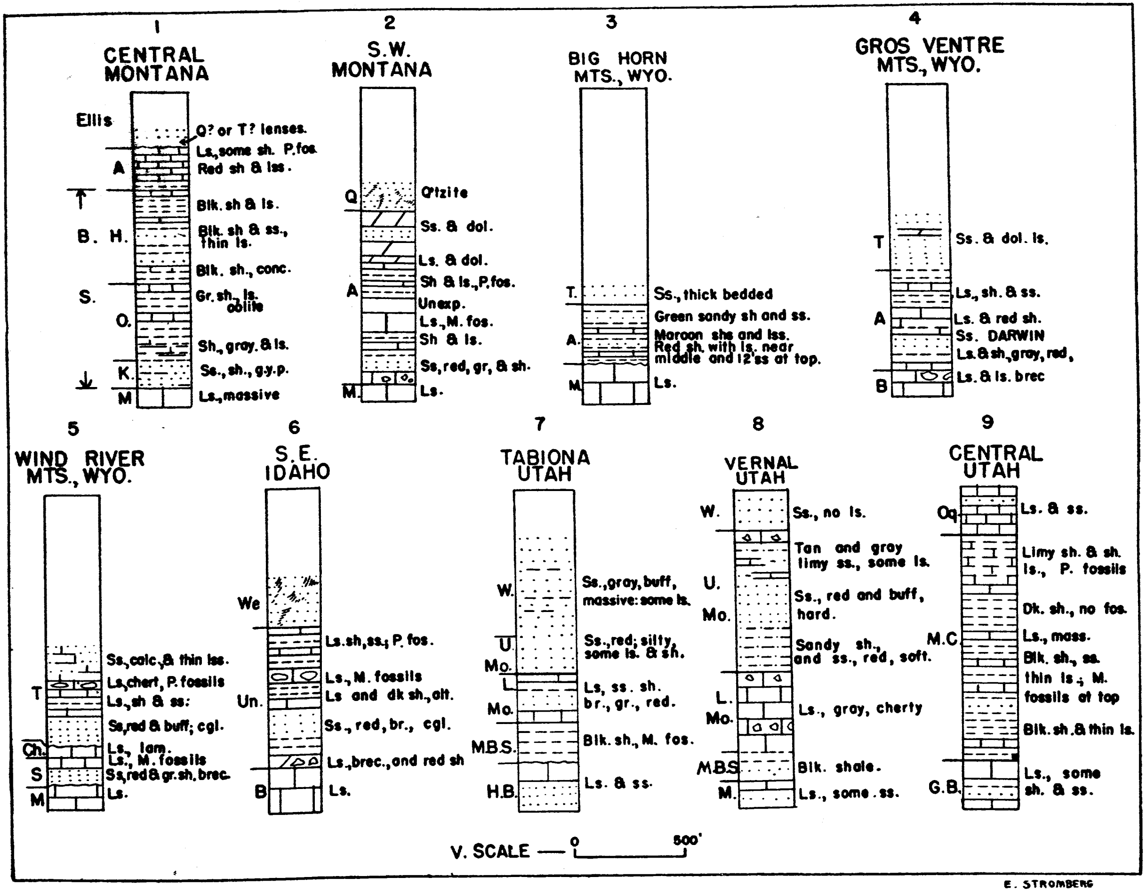



FIG. 2.—Selected stratigraphic sections crossing the Mississippian-Pennsylvanian boundary

Heath to be of formational rank and to belong to a group to which he applied the name Big Snowy group. Evidence from several classes of fossils published by different investigators has confirmed the Upper Mississippian age of the Big Snowy group and has suggested correlations of specific units with other units in the Mid-Continent region. The age of the Amsden of Scott of this region is not, however, firmly established. As stated earlier, Girty identified some of the faunules from it as upper Mississippian and others as Pennsylvanian. Scott in 1935 (p. 1032) considered the unit that he called Amsden in the Big Snowy Mountains to be of Mississippian age. Later in 1945 (1945*a*, p. 1195; 1945*b*, p. 1196), he assigned the Amsden of central Montana to the Pennsylvanian not only because, presumably, of a big overlap at its base that he had recognized in 1935 but also because, mainly, of the presence of certain species of *Millerella* in it. Other microfossils occur in the Amsden, but those so far identified are not definitely diagnostic. Though *Millerella* was for a time thought to be only of lowest Pennsylvanian age, it now is known to have a long range, which includes Mississippian as well as late Pennsylvanian, and some paleontologists maintain that species of *Millerella* are not yet safe zone markers. Perry and Sloss in 1943 (p. 1293) also described an overlap at the base of the Amsden, but they considered the Amsden to be both upper Mississippian and lower Pennsylvanian. The overlap, according to them, would then have occurred in upper Mississippian time. Thom and others have mentioned some evidence of physical unconformity at the base of the Amsden. Perry and Sloss also describe an unconformity at the base of the Big Snowy group, with sandstones of the Kibbey formation filling channels and solution cavities some 300 feet below the top of the Madison limestone. On the other hand, Scott (1935), Perry (1937), Pardee (1937), and others suggest an unconformity above the Amsden formation in Montana.

The three units that make up Scott's Big Snowy group have been mapped as formations in several places in Montana and have been recognized in subsurface as far east as the Dakotas and northward into Canada. The variations in thickness and paleogeography of these units are shown on maps by Perry and Sloss (1943). The Charles formation, proposed by Seager as part of the Big Snowy group below the Kibbey, has not yet gained wide recognition in surface work. The validity of Scott's use of the Kibbey, Heath, and Otter as formations would seem to the writer to be established either on the surface or in subsurface in those areas where each has sufficient thickness and distinction to qualify it as a mappable lithologic unit on the scales ordinarily used for topographic quadrangle

Explanation of Figure 2

The sections of figure 2 were selected to show interesting fossil occurrences, lithologic features, and stratigraphic terminology. Many of them are condensed in order that they may be shown together on one figure. The references are given in connection with the discussions of the sections on pp. 331–340. Only small parts of the lowest and uppermost formations in each of the sections are shown.

The following are the names for which the abbreviations in the respective sections stand: *1: B.S.*, Big Snowy group; *Q?*, Quadrant?; *T?*, Tensleep?; *A*, Amsden; *H*, Heath; *O*, Otter; *K*, Kibbey; *M*, Madison. *2: Q*, Quadrant; *A*, Amsden; *M*, Madison. *3: T*, Tensleep; *A*, Amsden; *M*, Madison. *4: T*, Tensleep; *A*, Amsden; *B*, Brazer. *5: T*, Tensleep; *Ch*, Chester; *S*, Sacajawea; *M*, Madison. *6: We*, Wells; *Un*, unnamed beds; *B*, Brazer. *7: W*, Weber; *U. Mo.*, Upper Morgan; *L. Mo*, Lower Morgan; *M.B.S.*, Mississippian black shale; *H.B.*, Humbug. *8: W.*, Weber; *U. Mo.*, Upper Morgan; *L. Mo.*, Lower Morgan; *M.B.S.*, Mississippian black shale; *M*, Mississippian. *9: Oq.*, Oquirrh; *M.C.*, Manning Canyon; *G.B.*, Great Blue.

work in the area in which it occurs. In other places these units would, if recognizable, be reduced to member or bed status. In this last alternative the question arises as to what formation they would be referred as members—an enlarged Amsden formation, a Big Snowy formation, or an entirely new formation. Terminology used in sections measured by Gardner, Hendricks, Hadley, and Rogers (1945) of the United States Geological Survey, suggests that the Kibbey, Otter, and Heath lose their identity, at least as formations, in short distances. This seems so because in some areas that are near the stratigraphic sections in which these authors recognize Scott's formations—Kibbey, Otter, and Heath—the same authors place all the beds between the Madison and the Quadrant in the Amsden. The Amsden as thus used may or may not contain equivalents of the Kibbey, Otter, and Heath. Some would maintain that these three units have pinched out and been overlapped by the Amsden. Another interpretation would be that the Kibbey, Otter, and Heath in some areas contain lenticular beds that change laterally so that the formations soon lose their identity in these areas. The apparent increase in thickness of the Amsden in places where the formations of the Big Snowy group are not recognizable and the Amsden is the only formation between the Madison and the Quadrant would favor the last interpretation. Unfortunately, fossils so far obtained are inadequate to solve this problem.

#### SECTION IN SOUTHWESTERN MONTANA

Section 2 of figure 2 is a section taken from those published by Gardner, Hendricks, Hadley, Rogers, and Sloss (1946), mentioned above. The section is condensed for graphic representation. The location is near Sappington, southwestern Montana, in Sappington Canyon, in Section 25, T. 1 N., R. 2 W., about 14 miles southwest of Three Forks and about 35–40 miles southwest of the Big Snowy Mountains. The formations shown are the Madison, the Amsden, and the Quadrant. A limestone breccia at the base of the Amsden formation suggests an unconformity. L. L. Sloss has identified Mississippian fossils, including *Diaphragmus*, from the lowest thick limestone unit shown in the Amsden of the section and Pennsylvanian fossils from a series of siltstones, 50–100 feet above the top of the limestone that bears Mississippian fossils. The top of the Amsden formation is here drawn at the top of the nonresistant beds, and a considerable thickness of alternating sandstones and dolomites is included in the Amsden. The uniformity in lithology of the Quadrant seems to have been the deciding factor in delineating it as a formation and in distinguishing it from the Amsden, which is recognized by its heterogeneity in composition and nonresistant character.

#### SECTION IN THE BIG HORN MOUNTAINS, WYOMING

Section 3 of figure 2, from the Big Horn Mountains, is condensed from one measured by Darton (1906, p. 5) in the canyon of Little Tongue River, Dayton quadrangle, Wyoming. This section was selected because it is only about 6 miles from the type locality of the Amsden formation. The type locality is given by Darton (1904, p. 396), who defined the formations as the Amsden Branch of Tongue River, about 5 miles southwest of Dayton, in the Dayton quadrangle, Wyoming. As shown in the figure, some of the thicknesses are slightly exaggerated. The total thickness of the Amsden is 190 feet. The terminology, which is that of

Darton, is widely used in this region at the present time. Judging from C. C. Branson's (1939, pp. 1202–1213) usage in sections in the Big Horn Mountains, probably some of the lower beds of Darton's section, possibly those up to the base of the 12-foot sandstone, would be included by him in his Sacajawea, or in other beds referred by him to the Mississippian. The typical area of Branson's Sacajawea formation, discussed later in this paper, is in the Wind River Mountains. To the writer's knowledge, Branson has not cited faunal evidence for extending it into the Big Horn Mountains, so presumably the extension was based mainly on lithology. No specific lithologic features, however, have been given to explain the extension or to furnish criteria for the lithologic differentiation of the Sacajawea from overlying beds.

In the Big Horn Mountains, the Amsden is a variable formation, both in thickness and in lithology. Because of the thickening and thinning of many beds in short distances, one suspects that at some stratigraphic zones they are lenticular. In several places the basal bed is a sandstone, which may attain a thickness of as much as 100 feet or more, but sandstones occur at several stratigraphic positions in the Amsden of most areas. In a section measured by the writer in 1920, along Little Goose Creek, about 25 miles southeast of the type locality, a thin-bedded sandy limestone that is dense to finely crystalline and has a purplish cast was considered the basal bed of the Amsden.

The contact with the underlying Madison limestone in this region in many places is irregular, and shales from the Amsden fill depressions some of which are probably sinkholes in the Madison.

The writer knows of no fossils from the type locality of the Amsden, but in print fossils have been reported from the Amsden of the Big Horn Mountains from several localities. The writer has collected them from other localities in the Big Horns. The presence of fossils of probable Carboniferous age from the upper part of the Amsden was mentioned by Darton (1904) in his original description of the formation. Pennsylvanian fossils from the Amsden have since been listed in many United States Geological Survey papers and in other publications. Some of these fossil lists were brought together by C. C. Branson in 1939. Fossils are not abundant in the lower part of the Amsden of the Big Horn Mountains; but, in 1906, Darton (p. 5) mentioned the occurrence of a coral identified by Girty as *Menophyllum excavatum* Girty and fragments of a *Spirifer* and a "Zaphrentis" from a limestone bed in the lower part of the Amsden near Soldier Creek, some 15–20 miles southwest of the type locality. Girty believed that these suggested Mississippian age for the lower Amsden; and the Amsden in this general area has for a long time been generally considered to contain both Mississippian and Pennsylvanian beds. The writer does not consider these fossils of themselves adequate to establish a Mississippian age for the lower part of the Amsden; but the absence of definite Pennsylvanian fossils near the base of the Amsden in the Big Horn Mountains and the presence of Mississippian fossils in the alternating series of beds between the Madison limestone and the Quadrant (Tensleep) formation north and west of the Big Horns in Montana and south and west of them in central Wyoming strengthen considerably the meager fossil evidence for a Mississippian age of the lower Amsden in the Big Horns. If definite stratigraphic tracing could show that the lowest beds of the Amsden of the Big Horns were

younger than the youngest near-by beds with Mississippian fossils, it would be an argument for the absence of Mississippian beds in the Big Horn Mountains, but this does not appear to be possible because of the areas of younger rocks between the mountain ranges. From time to time, verbal reports are made of the discovery of Pennsylvanian fossils in the basal Amsden of the Big Horns, but the writer does not know of any published record. Collections from the Amsden made in conjunction with field work in the Big Horns and mountains west of the Big Horn Basin by W. G. Pierce and D. A. Andrews and the writer have not yet been carefully studied.

In the Big Horn Mountains the sandstone near the base of the Amsden has been identified by some geologists as the Darwin sandstone member, whose type locality is in the Gros Ventre Mountains. This identification appears to be based on lithologic similarity and similar position in the stratigraphic succession. Some geologists would place the Mississippian-Pennsylvanian boundary at the base of the Darwin sandstone, not only in the Gros Ventres, but in other areas where they believe the Darwin can be recognized. Such an interpretation would not give a boundary closely tied to faunal data, as diagnostic fossils are not known from beds near this boundary. The practical advantages in mapping a unit boundary at the base of a prominent sandstone where that sandstone can be definitely identified are easily understood; but to assume without faunal data that such a boundary is a period or epoch boundary seems to the writer to be unjustified, even though an unconformity might be present. To use the base of the Darwin sandstone unit as the base of the Amsden, even if that sandstone can be definitely identified from section to section, would add still another type of lithologic criteria to those used elsewhere for the base of the Amsden, as there are several feet of nonresistant alternating shaly beds below the Darwin, or so-called "Darwin," at many places. The Amsden in many different places in the Big Horn Mountains and Big Horn Basin and at other places in Wyoming has been described and mapped by many geologists. Most geologists have considered it in part Mississippian and in part Pennsylvanian, but in local areas it has been referred entirely to the Pennsylvanian, and in a few areas, notably in the Wind River Mountains, it has been referred by a few geologists entirely to the Mississippian. Those who have placed the Amsden wholly in either the Mississippian or the Pennsylvanian have not cited definite fossil evidence from critical beds.

### SECTION IN THE GROS VENTRE MOUNTAINS, WYOMING

The section selected from the Gros Ventre area (sec. 4 in fig. 2) was measured by Wanless and Bachrach and published in 1947 by Helen L. Foster (p. 1557). The locality is north of Sheep Creek in the N.E. ¼, Section 10, T. 42 N., R. 115 W., Teton County, Wyoming. This locality is near the type area of the Darwin sandstone. Miss Foster designates the beds above the Darwin as upper Amsden; those below it as lower Amsden. The Darwin sandstone in this section is given as 97 feet thick, but its thickness varies in the surrounding region. The Amsden formation was examined briefly by the writer last summer at one locality in the Gros Ventres. The beds below the Darwin, in the lower part of the Amsden, consist of red shales, gray, pink, and lavender fine-grained limestones, and beds of chert. No fossil names were listed by Miss Foster, but she states that her "lower Amsden" contains fossils resembling both Mississip-

pian and Pennsylvanian types. She believes that the Mississippian-Pennsylvanian boundary is near the base of the Darwin. In connection with studies in this area by Eliot Blackwelder, C. W. Tomlinson, and other geologists of the United States Geological Survey in 1911 and in near-by areas in subsequent years, Dr. G. H. Girty examined many collections of fossils from beds referred to the Amsden. Lists from two different stratigraphic zones are published in Blackwelder's paper (1913, p. 176). The list for the uppermost zone, which is said to be from a thin group of limestone beds a little below the middle of the formation, shows the zone clearly to be Pennsylvanian. The fossils from a zone 60 feet lower are not so diagnostic but appear to the writer to be also Pennsylvanian. Girty, however, reserves the possibility that they may be Mississippian. There has not been an opportunity for the writer to re-examine the actual collections or to study Blackwelder's field notes. Neither collection is located stratigraphically with respect to the Darwin sandstone, but it is probable that both came from beds above it. This cannot, however, be definitely stated. The United States Geological Survey has had a field party working in the Gros Ventres during the past season, and several universities have had students and faculty members working there in recent summers. It is to be hoped that the examination of fossils obtained by these investigators will soon supply some definite faunal data on the age of the Darwin sandstone member of the Amsden.

#### SECTION IN THE WIND RIVER MOUNTAINS, WYOMING

Section 5, figure 2, shows relationships of the strata in the Wind River Mountains. The section was described by C. C. Branson in 1937 (p. 651), but the terminology is taken from a 1939 publication (pp. 1209–1210) showing the same beds. The locality is Bull Lake Creek. In his 1937 paper, Branson proposed to split the sequence that had previously been generally called Amsden into three units. The lowest of his three units consisted of 43 feet of cherty limestone underlain by 2–11 feet of "red and buff sandstone and shaly sandstone, breccia in places, shale cave filling in places." The 43-foot limestone is the zone of the invertebrate fauna described by E. B. Branson and D. K. Greger in 1918 (pp. 309–326) as of Ste. Genevieve age. For this lower unit of limestone and underlying red rocks Carl Branson proposed formation rank and gave them the name "Sacajawea formation." One of the reasons for giving these beds formation rank appears to have been that they were Mississippian, whereas the upper beds of the Amsden were Pennsylvanian. C. C. Branson considered the Sacajawea formation to range possibly from Salem to Ste. Genevieve age. He thus considered it pre-Chester. Beds that Branson probably would refer to the Sacajawea have been examined at several localities by the writer, but they contain very few fossils or none at all. Aside from the faunas described by E. B. Branson and Greger and by C. C. Branson in his 1937 paper, ostracodes have been described by Morey (1935, pp. 474–482) and by Croneis and Funkhauser (1938, pp. 331–360). These last two consider the ostracodes examined by them to be Chester.

Above the Sacajawea formation is a series of limestones about 60 feet thick that C. C. Branson believes is Mississippian, probably Chester, but few fossils have been found in it or, if found, have not been reported in publications. Above the limestone beds is a sandstone 80 feet thick that has been identified by some geologists as the Darwin sandstone. The

Mississippian-Pennsylvanian boundary is drawn at the base of this sandstone by some investigators; but, as elsewhere, there is little direct paleontologic evidence for its precise location at the base of the Darwin. Pennsylvanian fossils have been collected from a limestone some 80 feet above the top of the sandstone. Other geologists have rejected Branson's proposed Sacajawea formation on the grounds that in actual field practice it is not a mappable unit over an appreciable area. They have mapped the entire sequence of variable nonresistant beds between the Madison and Tensleep formations as the Amsden formation and have not ventured to indicate the precise location of the Mississippian-Pennsylvanian boundary within the Amsden. Some of these geologists have recognized the Darwin as a member of the Amsden. C. C. Branson, in 1939, proposed dropping the name Amsden. He would include in the Tensleep formation all those beds called Amsden by other investigators that are above the limestone beds referred by him to Chester? age. Geologists generally have not followed this suggestion, but it is used in figure 1 to illustrate various types of usage. The lower beds of the Amsden, as identified by most geologists, would be referred by Branson, as before stated, either to his Sacajawea formation or to the unnamed unit that he believes may be of Chester age.

### SECTION IN SOUTHWESTERN WYOMING

Section 6, figure 2, measured by W. W. Rubey and the writer along the Covey cutoff trail in the Salt River Mountains, near Afton, Wyoming, has been selected to show features present in that part of southeastern Idaho and southwestern Wyoming where the Mississippian-Pennsylvanian boundary has been drawn along the boundary between the Brazer and the Wells formations. The many difficulties accompanying the use of this boundary as a mapping boundary in the field led Rubey to erect a new mapping unit that includes the nonresistant and alternating thin beds that occur in both the lower part of the Wells and the upper part of the Brazer. A silty sandstone in the lower part of this unit may be the Darwin sandstone, or a lenticular sandstone other than the Darwin. Mississippian fossils, including *Diaphragmus elegans* (Norwood and Pratten) n. var., *Camarophoria* cf. *C. explanata* (McChesney), and *Linoproductus ovatus* (Hall), were collected from a zone in a massive limestone about 400 feet above the base of the nonresistant unit, and *Chaetetes milleporaceus* Milne-Edwards and Haime and other Pennsylvanian fossils, including a new species of *Orthotetina* that appears to be represented in the western United States only in Pennsylvanian or Permian rocks, from the beds immediately above the massive limestones. Specimens of a species of *Lithostrotionella* occur at the same zone as the *D. elegans* mentioned above. The lists of fossils identified in 1931 and 1932 from this section will be published in full in W. W. Rubey's report.

### MORGAN LIMESTONE AREA

The Morgan formation is typically exposed in Weber Canyon near the town of Morgan, Utah, where, in 1935, the writer measured a section including the Morgan, Brazer, and Weber formations (1936). The section has not been published because it seemed preferable to await detailed mapping to determine the extent and nature of the probable faulting. Girty had measured a section there some years before. The fossils collected during both investigations are being studied, but the studies have not been completed. A casual inspection of the col-

lections definitely confirms the statement of Blackwelder that the Morgan is, in part, of Pennsylvanian age. It has not been definitely determined whether Mississippian beds occur in the lower part of the Morgan. If the lower boundary of the Morgan is drawn on the basis of fossils and its age stated as Pennsylvanian, then, of course, Mississippian beds would be excluded from it. If, however, it is drawn on a lithologic basis that includes all beds in the nonresistant unit, then Mississippian beds might be present.

Since the writer's section was measured, A. J. Eardley (1944) has mapped in this area, and J. Stewart Williams (1943, p. 607) has published a description of the Morgan at the type area, where he found it to be 1,060 feet thick.

The Morgan formation has been extended south and east from the type area. Calkins and Butler (1943, p. 28) referred beds in the Cottonwood–American Fork area, Utah, to the Morgan(?). Others who have recognized the Morgan outside the type area include J. Stewart Williams (1943); Thomas, McCann, and Raman (1945); Huddle and McCann (1947*a*); McCann, Raman, and Henbest (1946); K. G. Brill, Jr. (1944); and Kinney and Rominger (1947).

The beds identified as Morgan by the above geologists vary considerably in lithology, and one might well ask whether it is advisable to carry the name Morgan so far afield. Thompson (1945, p. 31) has applied a new name, Hells Canyon formation, to beds in the Uintas that may be of Morgan age. Possibly his Younghall formation is equivalent to some part of the Morgan.

### SECTION NEAR TABIONA, UTAH

Section 7, figure 2, is taken from a paper by Huddle and McCann (1947*b*). At this locality the Morgan formation is divided on lithologic grounds into an upper and a lower unit. The contact of the formation with the overlying Weber is drawn at the place where the sandstones cease to be dominantly red. This coincides generally, but not precisely, with the beginning of the massive, coarser sandstones and the termination of soft silty and shaly sandstones. Red and purple sandstones and shales and red cherts occur in the dominantly nonresistant lower part of the Morgan. The Mississippian-Pennsylvanian boundary is placed at the top of a black shale from which the writer has identified fossils as Mississippian. Fossils from the lower part of the Morgan, above the black shale, that have so far been studied by the writer are either too incomplete or of types too generalized and long ranging to indicate a definite age determination, but they suggest Pennsylvanian age.

### SECTION NEAR VERNAL, UTAH

Section 8, figure 2, near Vernal, Utah, is condensed and generalized from a section, mostly along Ashley Creek, measured by D. M. Kinney and J. F. Rominger (1947) in the Whiterocks River–Ashley Creek area on the south flank of the Uinta Mountains, Utah.

As in the Tabiona region farther west, the Mississippian-Pennsylvanian boundary is placed at the top of a black shale, but the black shale is not so well exposed as in the Tabiona region. The Morgan is divided into two parts, both of which are tentatively considered to be of Pennsylvanian age. The collections need to be carefully studied, however, as they contain many forms that are generalized. The lower part of the Morgan is mostly limestone and is more resistant than at other localities where shales and sandstones are intercalated at short intervals. The upper part consists of three subdivisions, the lowest of which is mainly soft red shale, sandstone, or sandy shale.

It probably corresponds approximately to the upper unit of the Morgan in the area near Tabiona. The middle subdivision of the upper part of the Morgan is mainly hard buff to red sandstones, and the upper subdivision is tan limy sandstones and gray cherty limestones. It appears to the writer that the upper and middle subdivisions probably correspond to at least part of the beds considered to be Weber in the Tabiona region. The upper contact of the Morgan is drawn at the top of the highest limestone bed.

Stratigraphic sections here and in near-by parts of the Uintas have been made at or near the same localities as those already discussed by several geologists. The information in these separate investigations needs to be better integrated than it is at present, and work is now in progress to that end.

### SECTION IN STOCKTON-FAIRFIELD AREA, UTAH

Section 9 of figure 2, condensed from one given by Gilluly (1932, p. 31) from Soldier Canyon near Stockton, was selected to show features existing in an area where the Mississippian-Pennsylvanian boundary is within the Manning Canyon shale.

Both the upper and the lower contacts of the Manning Canyon shale are gradational. Gilluly places the upper contact, with the Oquirrh formation, at the point where the limestones start to become more abundant than shales. The lower contact is drawn where shales become more abundant than limestones. The Manning Canyon, in contrast to other formations farther north that include the Mississippian-Pennsylvanian boundary, is almost devoid of red beds. Mississippian (Chester) fossils occur about 500 feet above the base of the Manning Canyon, and Pennsylvanian fossils occur about 350 feet higher. These are listed in Gilluly's report.

The Manning Canyon shale has been recognized at various places in north-central Utah. A section showing the Manning Canyon shale in the Wasatch Mountains area near Provo is given by Baker (1947). Nolan (1935, p. 31) described the Manning Canyon of the Gold Hill area, central western Utah. Bissell and Hansen in 1935 (p. 163) discussed briefly the gradational character of the Mississippian-Pennsylvanian contact in Spring Creek Canyon, east of Springville, Utah.

## PROBLEMS IN PALEONTOLOGY

### NEED FOR MORE PALEONTOLOGICAL DATA

The need for more paleontologic data is definitely shown in the preceding discussions of the few selected stratigraphic sections. Not only is the need for additional and larger collections from certain zones indicated, but there is also shown a need for more studies and better integration of collections already made. The problem of additional collections from specific zones in critical areas is not everywhere easily solved. In many mountain ranges the Carboniferous rocks are exposed mainly in areas of high altitude that are difficult of access. Furthermore, many of the beds are either unfossiliferous or contain very few fossils. Some beds that have fossils do not yield them readily, and it is difficult to obtain specimens that are well enough preserved to permit definite identifications. When closely related genera and species differ from one another only in some small internal character or in some particular type of ornamentation, as many do, large collections are frequently required to provide specimens to show adequately these characters.

If one could everywhere determine the

age of a formation or parts of a formation by making collections from a few localities known to provide good material, the problem would be simplified. Collections definitely located geographically and stratigraphically from many widespread areas are, however, needed to determine the age ranges of formations. The determination of the stratigraphic positions of collections in the Rocky Mountain region is frequently a problem because of faulting and folding. Especially are so-called "bedding-plane faults" likely to be missed and duplications or omissions of strata unnoticed. Stratigraphic sections and the fossil collections from these sections, in the Rocky Mountain and Great Basin areas especially, are, as a rule, on much firmer ground if the stratigraphic work is done in connection with rather detailed mapping.

### SPECIES DIFFERENTIATION

Studies in systematic paleontology that deal with the relationships of the species of several genera need to be made. These should be made by paleontologists who have available large numbers of specimens, stratigraphically and geographically well located. Only by the study of large collections can differences between two individuals be correctly evaluated, that is, whether they are individual, varietal, specific, subgeneric, or generic. Extensive experience in applying zoölogical concepts in classification to closely related forms may substitute in part for the lack of large collections. The mere presence of some differences from a type specimen of a species does not, as all paleontologists know, constitute reasons for specific, or even varietal, differentiation. The ranges of variation within several species of the Rocky Mountain area need to be determined.

In the Rocky Mountain Carboniferous, certain individuals of the Pennsylvanian *Spirifer occidentalis-rockymontanus* group so closely resemble individuals of the Mississippian *S. increbescens-keokuk* group, as identified in the West, that one is forced to the conclusion that, if these species have been correctly identified, they either overlap or are not distinct species. Studies of western representatives of this group based on large, definitely located collections need to be made. The nomenclature should reflect as nearly as possible the actual relationships, and if in any instance it is desirable to combine two or more groups considered species into one, then that combination should be made. This would serve the cause of stratigraphic paleontology more than maintaining fictional differences would. Progress in paleontology, it is true, is made by finer and finer subdivisions, if they are truly useful, but progress is also made by combining so-called "species." It is frequently true that, when only a few specimens are known, several apparent species can be distinguished, whereas in larger collections species lines disappear and fewer species are recognizable.

Similar investigations need to be made of Mississippian and Pennsylvanian species of the genera *Composita*, *Chonetes*, *Lithostrotionella*, *Reticulariina*, *Linoproductus*, and others.

### USE OF FOSSILS IN CORRELATIONS

One of the most important problems in the use of fossils for correlation in the Carboniferous of the Rocky Mountain region has just been discussed. More needs to be known about the ranges in variation within certain common species and about effective criteria for distinguishing these species, if they are distinct. Another problem that has also been mentioned lies in the difficulty of obtain-

ing large collections of well-preserved materials at critical localities and at critical stratigraphic zones. Criticism has been made of lists of fossils that contain many question marks and provisional identifications, but species cannot be positively identified when the diagnostic characters are not preserved. Correlations can rarely be positively made if species or genera cannot be definitely identified.

Not all species in the Mississippian and Pennsylvanian rocks in the Rocky Mountain region are long ranging or generalized. Some are distinct and are of definite stratigraphic value, along with other evidence, in the correlation of beds and zones. They have been called "index fossils,"[2] but the writer hesitates to use this term because too many fossils are index fossils only as long as relatively little is known about them. Since the writer first began paleontological work in the West, the range of many so-called "index fossils" has been extended so that they can no longer be used in the restricted sense of the term. One of these is *Archimedes*. For many years it was considered an index of certain zones in the Mississippian in this country; but in 1927 Girty and Gilluly (1932) discovered it above *Chonetes mesolobus* (*Mesolobus mesolobus* of recent authors) in the Oquirrh Mountains, Utah; and it has since been discovered in many places in Pennsylvanian rocks in the West and in the Permian in Russia. Other fossils, to mention a few, that have lost their early meaning as index fossils, within the writer's experience, include *Lithostrotion* (or *Lithostrotionella*), *Caninia*, *Leptaena analoga* (Phillips), *Leiorhynchus carboniferum* Girty, *Reticulariina spinosa* (Norwood and Pratten), and *Millerella*.

Few paleontologists have ever considered correlations in the Rocky Mountain Carboniferous that were based solely on one or two index fossils to be more than temporary, nor have they postulated the presence or absence of beds or zones on a similar basis. In using index fossils, the possibility always exists that the ranges of these fossils may be extended, and it is preferable to use several types of fossil evidence, where possible, for every correlation. Whole faunas are decidedly more useful in making correlations than are conclusions based on several index fossils or on fossils of any one class.

The province of this paper does not permit an exhaustive discussion of the various uses of fossils in correlation or the theories behind these uses. Paleontologists working on late Paleozoic rocks in the Rocky Mountains are generally aware of the many complex problems in any correlation and are usually on the alert for influences of less obvious factors. For instance, in parts of the Rocky Mountain region certain types or classes of fossils are relatively common in specific types of lithology and nearly absent in others—to cite one example, the large horn corals of the crystalline limestones of the Brazer. The possibility is always kept in mind that these corals are absent at certain stratigraphic zones merely because of ecological conditions (to be discussed more fully later). Another example: certain types or classes of fossils, such as crinoids and cephalopods, are relatively rare, and it is possible that their assumed excellence as zone markers is enhanced by their rareness. Still another example: several Mid-Continent

[2] An "index fossil" is generally defined as one that is characteristic of a specified time unit, more or less irrespective of facies, whereas a "facies fossil" is defined as being characteristic of a certain facies and crossing one or more time boundaries. In the larger aspect, of course, all fossils are index fossils and all fossils are probably facies fossils, but most geologists define and use the terms in the more restricted sense.

species have been identified in the Rocky Mountains, and it is possible that some of these species migrated into the Rocky Mountains so slowly that they occur there at a later time than in the Mid-Continent region. The possibility is kept in mind that the absences of species from certain zones may mean only inadequate collecting; also that classes of fossils only recently recognized and as yet little known may yield results that appear to be more definite than if they were better known. The value of relative abundance of specific forms in making local correlations has for a long time been realized and utilized in the Western states.

Not only do paleontologists use comparisons of total faunas and combinations of index fossils for correlations, but the reported first appearances of new forms and the reported extinctions of old forms are sometimes given special consideration. Certain evolutionary stages in specific species and especially genera, if adequately tested, and certain trends in evolution, if firmly established, are also used. An example of this last is the tendency for coarsely plicated Spirifers of the *rockymontanus* type to develop prominently in the middle Mississippian and to die out at the top of the lower Pennsylvanian (Des Moines equivalent). It is realized that whole faunas may transgress time lines, but such transgressions are relatively unimportant in correlations within local areas.

## ZONATION

For many years rocks of Chester age have been generally separated from rocks of older Mississippian age throughout the northern and central Rocky Mountain area, and certain Mid-Continent zones have been recognized within the pre-Chester Mississippian rocks in several local areas. As early as 1873, Meek (p. 433) published a notice of a fauna from beds that he assigned to the Spergen limestone, which was then considered part of the St. Louis limestone, and that are probably now included in the Madison limestone in Montana. The presence of a zone of St. Louis or Meramec age in the Mississippian rocks of the Rocky Mountains has been mentioned in print by Willis (pp. 316, 324) in 1902 (St. Louis fossils identified by Stuart Weller in the Yakinikak limestone of Willis in northwestern Montana); by Girty (1927, p. 71), both published and unpublished; by Sloss and Laird in 1945; and by several others. The writer and others have recognized it in the field at several places where its presence has not been mentioned in published accounts. Other zones in the Mississippian that correspond to pre-Chester Mississippian zones in the Mid-Continent area have been recognized in local areas.

In 1927, in a general summary of the Mississippian rocks of southeastern Idaho, Girty (p. 71) recognized affinities of different faunules from the Brazer with those from the Spergen, St. Louis, and Chester strata. He realized, however, that the Spergen fauna may be a facies fauna rather than one that can be definitely correlated with the typical Spergen. The writer in 1935 and also in later years measured and collected from many stratigraphic sections in an effort to fix these and possibly other zones in the Brazer, as well as to determine possible zones in the Wells and Phosphoria formations. J. Stewart Williams and J. S. Yolton in 1945 recognized five zones in the Brazer limestone as exposed at Dry Lake, Utah. The lowest zone was correlated by them with the Warsaw formation and the next highest zone with higher beds in the Meramec group, possibly with the St. Louis limestone. Their

third highest zone is considered by them to be transitional between the Meramec and Chester groups, and the two upper zones are Chester. Most of the Chester forms cited by them are generalized forms that have not only a considerable stratigraphic range in the Chester but occur or are represented by very closely related forms both in younger and in older rocks. One exception, however, is *Diaphragmus elegans* (Norwood and Pratten). The proposed correlation of Branson's Sacajawea formation (lower Amsden of most authors) with Mid-Continent beds and of the units of the Big Snowy group with specific units in the Chester of the Mid-Continent has been discussed before.

On the Pennsylvanian side of the Mississippian-Pennsylvanian boundary, a zone of post-Morrow–pre-Kansas City age has been recognized widely for a long time. The zone was first recognized and identified widely by megafaunas (Williams, in Moore, 1944); but fusulinids attesting to its existence were also identified by Girty prior to 1927. During recent years fusulinids from the Quadrant formation of Montana, the upper part of the Amsden and Tensleep of Wyoming, the Wells formation of Idaho and Utah, the Oquirrh of Utah, and other formations in neighboring states have supplemented and confirmed the evidence from the larger invertebrates. Although some efforts have been made to divide this zone into two—a lower Lampasas and an upper Des Moines—it has not proved feasible on the basis of the megafossils. Some fusulinid workers also have not found such a division practicable.

Correlation of Lower Pennsylvanian beds with the Pennsylvanian formations of the Appalachian region rather than directly with the formations of the Mid-Continent areas was attempted by G. H. Girty in several written opinions, and he has identified, more or less provisionally, beds of Pottsville and post-Pottsville age. The writer does not believe that these correlations, when based on invertebrate megafossils, were very reliable as to detail, and Girty appears not to have considered them very reliable. In 1934, C. B. Read described a Pottsville flora, probably of middle Pottsville age, from beds exposed near Leadville, Colorado, from which Girty had described a macrofauna that he believed represented a zone that was "very early in Pennsylvanian time and probably older than any beds of the Kansas and Nebraska section." Read regarded this flora as older than the Glen Eyrie flora.

During the last five or six years, beds thought to be immediately above the Mississippian-Pennsylvanian boundary have been referred to a zone that is considered the equivalent of the Morrow of the Mid-Continent. Most of these references have been based mainly on fusulinid evidence. In general, the larger invertebrate fossils, where they are present, do not give adequate evidence—if, in fact, any positive evidence at all—for distinguishing the so-called "Morrow zone" from the zones bearing both Des Moines–Lampasas macrofossils and microfossils. This does not necessarily mean, however, that larger collections may not provide material whereby such a zone can be recognized by larger fossils.

L. G. Henbest (1946) seems to have been the first actually to designate beds in the area under discussion as of Morrow age, but the presence of very early Pennsylvanian beds in near-by areas had been noted before by C. B. Read, G. H. Girty, the writer (1944, p. 700), and others, and M. L. Thompson (1936) had previously suggested the presence of beds of Morrow age in the Black Hills. Paleon-

tologists had also identified a zone principally in the subsurface that they called "Morrow" in other parts of the general Rocky Mountain area, namely, New Mexico and Colorado. M. L. Thompson (1945) identified a zone correlated by him with the Morrow over a wide area in the Uinta Mountains; Williams and Yolton (1945, p. 1152) referred rocks in central-northern Utah to the Morrow; Thomas, McCann, and Raman (1945) referred rocks in northwestern Colorado and northeastern Utah to the Morrow; and H. W. Scott (1945, p. 1195) referred rocks in the sequence which he had identified as Amsden in central Montana to the Morrow.

In assigning rocks to the Morrow, Henbest appears to have laid considerable stress on the occurrence in them of the fusulinid genus *Millerella*, which was once thought to be restricted to rocks of Morrow age; but the range of *Millerella* is now generally considered to extend from late Mississippian to late Pennsylvanian time. Thompson recognizes the long range of the genus *Millerella* in Pennsylvanian rocks and the difficulty of using species of *Millerella* to denote different stratigraphic zones. He identifies his zone of *Millerella*, which zone name he applies to rocks of Morrow age, by the predominance of specimens of *Millerella* and the absence of the more highly developed fusulinids with which *Millerella* is commonly associated in Pennsylvanian rocks of post-Morrow age. Williams and Yolton base their assignment of beds to a Morrow age mainly on microscopic bryozoans and ostracodes identified by Chalmer L. Cooper. They also list larger fossils, but nearly all the larger fossils are forms which, if identifiable, have been collected elsewhere in beds associated with fusulinids of younger age than Morrow. Scott bases his age determination on *Millerella*, with supplementary evidence from other microscopic forms.

Despite the recognition of beds at the places mentioned above and at other places in the northern and central Rocky Mountains, correlated with faunal zones in the Mid-Continent, zonation studies useful over wide areas in the rocks near the Mississippian-Pennsylvanian boundary have not been published. One reason for this deficiency is the paucity of fossils and the poor preservation in many beds. Large collections of good fossils from many zones and localities are needed if zonation is to be based on trustworthy grounds. Another reason is the great number of relatively generalized long-ranging species. Still another reason is that faunal zones do not coincide in many instances with lithologic divisions, as they more or less do in some places. This reduces the general use and recognition of faunal zones in areas where the natural mapping units are not also faunal units.

In the Rocky Mountain area there are great thicknesses of rather uniform lithology, especially in the Mississippian rocks. This contrasts with the situation in the Mississippi Valley, where comparatively frequent alternations in deposition have occurred and where, consequently, the typical formations are thinner lithologic units. The Mississippian faunas in the Mississippi Valley were thus subjected to more frequent changes of conditions than were those in the Rocky Mountain area. Both the Mississippian and the Pennsylvanian rocks of the Mid-Continent were deposited in a different basin (though there were connections) than the one in which the rocks of the same approximate ages in the Rocky Mountain region were deposited. Although some species have been identified as Mid-Continent species, the faunas

in general are quite different. Most correlations of zones in the Rocky Mountains with zones in the Mid-Continent, when made in detail, have appeared based upon inadequate evidence. The writer has believed for a long time that either the southeastern Idaho section or one of the sections in central Utah in an area that has been mapped in detail and from which a great many collections have been made should be selected as a Rocky Mountain standard and that correlations in this region should be made in terms of this standard rather than in terms of Mid-Continent standards. There seems no reason to suppose that in this different basin the ranges of fossils would be precisely the same as in the Mid-Continent basin or basins. Broad correlations can be made, but only broad ones are justified at the present. Several instances are on record of ranges of Mid-Continent species, and even of genera, that are different from ranges in the Mid-Continent.

## STRATIGRAPHIC PROBLEMS

### CRITERIA FOR FORMATIONS

A study of the selected stratigraphic sections discussed on pages 331–340 shows that a considerable variety of criteria has been used for the definition and delimitation of formations near the Mississippian-Pennsylvanian boundary in the northern and central Rocky Mountain region. These criteria have given quite different results concerning the content and thickness of a formation in the same or in near-by regions. Most geologists have used lithologic criteria, but many have combined lithologic criteria with structural relations, such as unconformities, karst surfaces, or brecciation, or with faunal data. In some instances the location of an unconformity or the boundary of a faunal zone seems to have been the determining factor in the separation of one formation from another, and the lithologic change at the boundary only secondary. Few geologists definitely state that the basis for their formation boundaries is faunal, but those who refuse to put Mississippian beds in the same formation with Pennsylvanian beds, simply because of this difference in age, are nevertheless using faunal data as their main criterion.

Even those who have used lithologic composition as the main factor in the definition of their formations have not agreed on the type of lithologic criteria to use. It is probably unlikely that absolute uniformity in usage will ever be attained over wide areas or in different sections of the country; and it probably is not desirable. Geologists as a rule will use whatever they find in the particular region that provides mappable units, and the choice will depend not only on individual preferences but also on the character of the topography and climate of the particular area, the lithology of the rocks, and other factors.

In work in the Rocky Mountain region, the writer has been impressed with the usefulness of lithologic formations based upon features of the total lithologic composition of the formation more than that of formations based on the presence of some specific color; or on the highest or lowest occurrence of some particular type of lithology, such as the highest limestone or the lowest phosphate bed; or on the tracing of some conspicuous bed, such as a sandstone; or on the tracing of unconformities, either by faunal data or by matching unconformities seen in separate outcrops. Nearly all these criteria have been, or are being, used in the Rocky Mountain area, and local conditions might make any one of them more desirable than the others. The first ap-

pearance of a changed type of lithology especially has cogent arguments in its support.

Far more important than the adoption of any one particular type of criterion is the desirability of uniform usage, at least in local areas, to which, after all, formation names only apply.

During recent years detailed stratigraphic work and subsurface work have greatly increased in the Rocky Mountain area. This work has added much needed detailed knowledge regarding the individual beds that make up the formations and has resulted, in some areas, in the breaking-up of larger formations and the giving of formational rank to units that constituted subdivisions or merely unrecognized parts of the larger formations. In the writer's opinion, new and thinner formations are desirable if, and wherever, they conform to the generally accepted criterion that a formation "shall . . . meet the practical and scientific needs of the users of geologic maps" (Ashley *et al.*, 1933, p. 431). As stated in the remarks accompanying the stratigraphic code just quoted, "practicability of mapping is usually an essential feature" of a formation. It is only "under exceptional conditions" that this criterion should be waived. The test of the validity of new formations proposed for use in rocks of Carboniferous age in the Rocky Mountain area, whether surface or subsurface formations, would, in the writer's opinion, lie in the decision as to whether or not they could reasonably be supposed to form practical mapping units, under the conditions existing in the region at the present time, if exposed over a considerable area. The validity of any surface formation can be tested by mapping. Ample terminology exists for the recognition and designation of lithologic units below the rank of formation, so that distinctions as close as are desired can be recognized by some sort of designation. As indicated in the remarks explaining features of the code, formations may include parts of two systems or even major unconformities where there is a sequence of similar beds and where in practical work it is not useful to make a division.

Much has been written about the necessity of distinguishing between rock and time (or so-called "time") units in stratigraphic nomenclature, and interest in this question has been revived lately because of many examples of widespread confusion and because a third category—time-rock units—has been proposed. It is not appropriate to discuss this question here, but a satisfactory understanding of these differences, if agreed upon, would aid in achieving more uniformity in usage in some parts of the Rocky Mountains.

### UNCONFORMITIES

Unconformities exist between and within several formations that occur near the Mississippian-Pennsylvanian boundary in the northern and central Rocky Mountains. Many of these are discussed in connection with the different stratigraphic sections, to which the reader is referred, but additional data on many of the unconformities are in publications from which no sections were selected or in the writer's files and notes.

Williams and Yolton (1945, p. 1150) show a widespread unconformity at the base of the Meramec division. Unconformities in local areas at this stratigraphic zone have been mentioned by others. A widespread unconformity occurs at the top of the Madison limestone over a wide area. The writer believes that this is mainly a pre-Chester or pre–Ste. Genevieve unconformity. In some places in Montana and Wyoming, beds that have been considered as young as St.

Louis in age have been included in the Madison; but in other places the upper part of the Madison limestone is of Osage age. The disagreement regarding the age of the Madison is beyond the scope of the present paper. Some of the disagreement may be due to a difference in interpretation of the boundaries of formations and groups in the Mississippi Valley, with which correlations have been attempted. An unconformity and overlap between the Heath and the Amsden formations that may be within the Chester has been described from Montana, but the age of the unconformity depends on the age assigned to the Amsden there. An unconformity occurs beneath the Darwin sandstone unit at some places where it has been identified in Wyoming. Unconformities have been placed at the Mississippian-Pennsylvanian boundary as delimited by fossils mainly because it appears that in some places fossil zones of the Mississippi Valley Basin are not represented here. Lithologic criteria that suggest terrestrial beds and other types of lithologic criteria indicate unconformities at various places in the Amsden formation, some of which are probably very local in extent.

## PROBLEMS IN ECOLOGY AND PALEOGEOGRAPHY

Problems in ecology and paleogeography that are as challenging as those in paleontology and stratigraphy occur in the Carboniferous rocks of the northern and central Rocky Mountains. Mention may be made of a few. The occurrence in these rocks of black shales, dwarf faunas, oölitic limestones, great thicknesses of crystalline limestones containing few fossils other than corals, alternating terrestrial and marine beds, considerable thicknesses of red beds, finely laminated limestones, gypsums, and cross-bedded quartzites suggests the wide range of ecological conditions that existed during late Mississippian and early Pennsylvanian time. These also indicate the variety of ecological problems.

Existing data on the paleogeography need to be consolidated and published and new data added. Geosynclinal and shelf areas have been indicated in parts of the area. The location of major land masses is known in a general way, but details of the paleogeography, including precise locations and extent of major features, are not known. Relationships to other areas, such as the Mid-Continent and Pacific Coast areas, need to be more definitely worked out and sea connections more precisely located. Much additional data are needed on sources of sediments. Additional isopach maps need to be constructed; but these, to be widely understood, either will have to await unification of some of the stratigraphic terminology or will have to be accompanied by considerable stratigraphy, explaining the author's usages of formations.

Of quite a number of recent articles containing maps or discussions of paleogeographic features, papers by Nolan (1943) and Eardley (1947) are especially comprehensive and significant and contain valuable references to other literature.

## CONCLUSIONS

1. As elsewhere in the United States, many unsolved problems are presented by rocks at and near the Mississippian-Pennsylvanian boundary in the northern and central Rocky Mountains.

2. The problems cover various phases of stratigraphy, systematic paleontology, stratigraphic paleontology, ecology, and paleogeography.

3. Adequate collections of well-pre-

served fossils are needed from critical stratigraphic zones at critical localities. Collecting has been limited in some places by the absence of marine invertebrates due to unfavorable conditions either during the deposition of the beds or for the preservation as fossils. At some other places it is limited by the difficulty of separating fossils from the enclosing rock. At many places, however, the small collections made have shown that more time and more experience in collecting would have yielded the larger and more varied collections needed.

4. The considerable amount of paleontologic data already obtained should be better organized for effective use.

5. Critical paleontologic studies of several common species are essential. These should be based on enough specimens to show the range of variation within each species. Species that are too narrow because of the lack of knowledge of the range of individual variation encourage erroneous and unsubstantiated correlations. Species that are too broad contribute to the difficulty of establishing recognizable paleontologic zones. The fact that certain species definitely are long ranging and do not contribute effectively to detailed zonation studies should be recognized and the nomenclature adjusted accordingly.

6. Paleontologic correlations should be based on several lines of evidence. Index fossils are very useful in combination with one another and with other faunal data, but changes in the stratigraphic ranges of some so-called "index fossils" argue for caution in the use of all of them. Correlation data from one class of fossils need to be checked against those from other classes wherever possible. No one class is of itself totally sufficient.

7. Correlations of restricted paleontologic zones in the Rocky Mountains with zones in the Mid-Continent region based on evidence from larger invertebrate fossils have in several instances been made on data that are inadequate in both places.

8. Detailed correlations within the Rocky Mountain area should be made mainly with typical sections within that area. There is little reason to believe that the Rocky Mountain and Mid-Continent areas, being distinct, though in places connected, basins, had the same geologic history. Nor did they necessarily have the same faunal zones, if one uses the term "zone" in a restricted sense, or precisely the same ranges of species. The gross differences in the faunas suggest that the ranges of species may, in fact, be quite different.

9. The use of fossils in determining the relative ages of beds in the Carboniferous of the Rocky Mountain area has been extensive and, despite the deficiencies of the data at some places, appear to have been as successful, though perhaps not so detailed, as in most other areas. Many hundreds of collections have been studied, and the age conclusions based upon these studies have been given to field men for further work. Many hundreds of square miles involving Mississippian and Pennsylvanian rocks have been mapped in detail, and some areas have been remapped on different scales. Many areas have been complexly folded and faulted, and other areas have been highly metamorphosed. Yet detailed remapping of some areas and mapping of neighboring areas have changed few of the paleontological age determinations.

10. Formations have been defined and their limits set by many different criteria. More uniformity than now exists, as to both criteria and terminology, is desirable. A big step toward uniformity would result from general agreement to

define and actually employ formations, whether surface or subsurface, as lithologic units which could be effectively mapped, or could logically be supposed to be mappable, on the ordinary scales of topographic quadrangle mapping used in the area where they occur. Smaller units could be given varying degrees of subformational rank to allow as fine discrimination as desired. Such a definition would not prohibit formations from containing unconformities, parts of two geologic systems, or several paleontologic zones or from varying somewhat in age from place to place. Even with such a general agreement, problems regarding the details of the lithologic composition of various formations, lateral gradation, and other points would remain to be solved.

11. Additional work on the Carboniferous rocks and their faunas in the Rocky Mountain area should be very thorough and detailed, or frankly interpretive in localities for which the data are inadequate. All types of previous knowledge should be utilized. Reconnaissance work has been done in nearly all areas.

12. The combination of detailed stratigraphic and paleontologic work with detailed mapping offers the best possibility of obtaining adequate and reliable data needed from the northern and central Rocky Mountain region.

## REFERENCES CITED


Ashley, G. H., *et al* (1933) Classification and nomenclature of rock units: Geol. Soc. America Bull. 44, pp. 423–459.

Baker, A. A. (1947) Stratigraphy of the Wasatch Mountains in the vicinity of Provo, Utah: U.S. Geol. Survey Oil and Gas Inv. Preliminary Chart 30.

Bissell, H. J., and Hansen, G. H. (1935) The Mississippian-Pennsylvanian contact in the central Wasatch Mountains, Utah: Utah Acad. Sci. Proc., vol. 12, p. 163.

Blackwelder, Eliot (1910) New light on the geology of the Wasatch Mountains, Utah: Geol. Soc. America Bull. 21, pp. 517–542.

——— (1913) New or little known Paleozoic faunas from Wyoming and Idaho: Am. Jour. Sci., 4th ser., vol. 36, pp. 174–179.

Branson, C. C. (1937) Stratigraphy and fauna of the Sacajawea formation, Mississippian, of Wyoming: Jour. Paleontology, vol. 11, pp. 650–660.

——— (1939) Pennsylvanian formations of central Wyoming: Geol. Soc. America Bull. 50, pp. 1199–1225.

Branson, E. B., and Greger, D. K. (1918) Amsden formation of the east slope of the Wind River Mountains of Wyoming and its fauna: Geol. Soc. America Bull. 29, pp. 309–326.

Brill, K. G., Jr. (1944) Late Paleozoic stratigraphy, west-central and northwestern Colorado: Geol. Soc. America Bull. 55, pp. 621–655.

Calkins, F. C., and Butler, B. S. (1943) Geology and ore deposits of the Cottonwood–American Fork area, Utah: U.S. Geol. Survey Prof. Paper 201.

Croneis, C. G., and Funkhauser, H. J. (1938) New ostracodes from the Clore formation [Ill.]: Denison Univ. Bull. 38, no. 10 (Sci. Lab. Jour., vol. 33, art. 7), pp. 331–360.

Darton, N. H. (1904) Comparison of the stratigraphy of the Black Hills, Big Horn Mountains and Rocky Mountain front range: Geol. Soc. America Bull. 15, pp. 379–448.

——— (1906) Description of the Bald Mountain and Dayton quadrangles, Wyoming: U.S. Geol. Survey Geol. Atlas, Bald Mountain–Dayton Folio 141.

Eardley, A. J. (1944) Geology of the north-central Wasatch Mountains, Utah: Geol. Soc. America Bull. 55, pp. 819–894.

——— (1947) Paleozoic Cordilleran geosyncline and related orogeny: Jour. Geology, vol. 55, pp. 309–342.

Foster, H. L. (1947) Paleozoic and Mesozoic stratigraphy of northern Gros Ventre Mountains and Mount Leidy Highlands, Teton County, Wyoming: Am. Assoc. Petroleum Geologists Bull. 31, pp. 1537–1593.

Gardner, L. S.; Hendricks, T. A.; Hadley, H. D.; and Rogers, C. P.; Jr. (1945) Columnar sections of Mesozoic and Paleozoic rocks in the mountains of south-central Montana: U.S. Geol. Survey Oil and Gas Inv. Preliminary Chart 18 (also published as Stratigraphic sections of Upper Paleozoic and Mesozoic rocks in south-central Montana, with description of fauna of Amsden and Heath formations by L. L. Sloss: Montana Bur. Mines and Geology Mem. 24, 1946).

GILLULY, JAMES (1932) Geology and ore deposits of the Stockton and Fairfield quadrangles, Utah: U.S. Geol. Survey Prof. Paper 173.

GIRTY, G. H. (1927) In MANSFIELD, G. R., Geography, geology, and mineral resources of part of southeastern Idaho: U.S. Geol. Survey Prof. Paper 152.

HENBEST, L. G. (1946) Correlation of the marine Pennsylvanian rocks of northern New Mexico and western Colorado (abstr.): Washington Acad. Sci. Jour., vol. 36, p. 134.

HUDDLE, J. W., and McCANN, F. T. (1947*a*) Geologic map of Duchesne River area, Wasatch and Duchesne counties, Utah: U.S. Geol. Survey Oil and Gas Inv. Preliminary Map 75.

——— — ——— (1947*b*) Late Paleozoic rocks exposed in the Duchesne River area, Duchesne County, Utah: U.S. Geol. Survey Circ. 16.

KINNEY, D. M., and ROMINGER, J. F. (1947) Geology of the Whiterocks River–Ashley Creek area, Uintah County, Utah: U.S. Geol. Survey Oil and Gas Inv. Preliminary Map 82.

McCANN, F. T.; RAMAN, N. D.; and HENBEST, L. G. (1946) Section of Morgan formation, Pennsylvanian, at Split Mountain in Dinosaur National Monument, Uintah County, Utah: U.S. Geol. Survey Mimeographed Rept., 1946.

MEEK, F. B. (1873) Preliminary paleontological report, consisting of lists and descriptions of fossils, with remarks on the ages of the rocks in which they were found: U.S. Geol. and Geog. Survey Terr. 6th Ann. Rept., pp. 429–518.

MOREY, P. S. (1935) Ostracoda from the Amsden formation of Wyoming: Jour. Paleontology, vol. 9, pp. 474–482.

NOLAN, T. B. (1935) The Gold Hill mining district, Utah: U.S. Geol. Survey Prof. Paper 177.

——— (1943) The Basin and Range Province in Utah, Nevada, and California: U.S. Geol. Survey Prof. Paper 197-D, pp. 141–196.

PERRY, E. S., and SLOSS, L. L. (1943) Big Snowy group, lithology and correlation in the northern Great Plains: Am. Assoc. Petroleum Geologists Bull. 27, pp. 1287–1304.

READ, C. B. (1934) A flora of Pottsville age from the Mosquito Range, Colorado: U.S. Geol. Survey Prof. Paper 185-D, pp. 79–96.

REEVES, FRANK (1931) Geology of the Big Snowy Mountains, Montana: U.S. Geol. Survey Prof. Paper 165, pp. 135–149.

SCOTT, H. W. (1935) Some Carboniferous stratigraphy in Montana and northwestern Wyoming: Jour. Geology, vol. 43, pp. 1011–1032.

——— (1945*a*) Age of the Amsden formation (abstr.): Geol. Soc. America Bull. 56, p. 1196.

——— (1945*b*) Pennsylvanian stratigraphy in Montana and northern Wyoming (abstr.): *ibid.*, p. 1196.

SLOSS, L. L. and LAIRD, W. M. (1945) Mississippian and Devonian stratigraphy of northwestern Montana: U.S. Geol. Survey Oil and Gas Inv. Preliminary Chart 15.

THOMAS, C. R.; McCANN, F. T.; and RAMAN, N. D. (1945) Mesozoic and Paleozoic stratigraphy in northwestern Colorado and northeastern Utah: U.S. Geol. Survey Oil and Gas Inv. Preliminary Chart 16.

THOMPSON, M. L. (1936) Fusulinids from the Black Hills and adjacent areas in Wyoming: Jour. Paleontology, vol. 10, pp. 95–113.

——— (1945) Pennsylvanian rocks and fusulinids of east Utah and northeast Colorado correlated with Kansas section: Kansas Univ. Geol. Survey Bull. 60.

WILLIAMS, JAMES STEELE (1936) Stratigraphic section and faunules of some western Carboniferous formations at or near the type localities (abstr.): Geol. Soc. America Proc. 1935, pp. 118–119.

——— (1944) In MOORE, R. C., *et al.*, Correlation of Pennsylvanian formations of North America: Geol. Soc. America Bull. 55.

WILLIAMS, JAMES STEWART (1943) Carboniferous formations of the Uinta and northern Wasatch Mountains, Utah: Geol. Soc. America Bull. 54, pp. 591–624.

——— and YOLTON, J. S. (1945) Brazer (Missisippian) and lower Wells (Pennsylvanian) section at Dry Lake, Logan quadrangle, Utah: Am. Assoc. Petroleum Geologists Bull. 29, pp. 1143–1155.

WILLIS, BAILEY (1902) Stratigraphy and structure, Lewis and Livingston Ranges, Montana: Geol. Soc. America Bull. 13, pp. 305–352.

# AMERICAN MISSISSIPPIAN AMMONOID ZONES

A. K. MILLER
State University of Iowa, Iowa City, Iowa

## ABSTRACT

No comprehensive study of our Carboniferous ammonoids has been attempted since the very beginning of the present century. Only three ammonoid zones can now be recognized in the American Mississippian system. These seem to be more or less equivalent to our Kinderhook, Osage, and Meramec-Chester beds, and they are characterized by the genera *Protocanites*, *Beyrichoceras*, and *Goniatites*, respectively. Locally, ammonoids are abundant in the Kinderhook and the Meramec, but elsewhere they are rare. Our Devonian and Mississippian systems are not known to carry a single ammonoid genus in common, and only two stocks cross the border between them. One of these, the prolecanitids, became quite varied in the Mississippian, and they have a considerable amount of stratigraphic value there. The other stock that continues into the Mississippian is represented by the genus *Imitoceras*. Its descendants underwent a great development during the Carboniferous and gave rise to most of the many diverse forms known from there. Although Kinderhook ammonoids differ materially from those of the Devonian, they are close to those of the Osage. The latter, known from only a few specimens, are quite distinct from Meramec-Chester forms, which are more or less transitional with early Pennsylvanian types.

# KINDERHOOK MICROPALEONTOLOGY[1]

CHALMER L. COOPER
United States Geological Survey

ABSTRACT

Both published and unpublished information on Kinderhook microfaunas, together with significant macrofaunas and floras, are reviewed in an effort to assess the value of these microfaunas, particularly the conodonts, for stratigraphic correlation. Of all classes of fossils represented, the conodonts are best known because of their widespread occurrence in beds of this age and the relatively large number of papers published on them. The ostracodes, which occur in several zones, are relatively abundant but little known. Other kinds of microfossils, such as Foraminifera, holothurian and echinoid fragments, spores, ?radiolarians, and nepionic forms of many of the common larger fossils are the subject of few reports or have been neglected entirely.

## INTRODUCTION

Any faunal or floral analysis of the Kinderhook, whether micro or macro, must take into consideration the whole biota, together with its stratigraphic relationships. Because the Kinderhook is the lowest or oldest of the Mississippian series, its faunas must be separated from those of the Upper Devonian, and the plane of division between the strata of the two systems must be located. There are numerous approaches to this problem. Because various methods have been used by workers during the past century, many and varied conclusions have been reached. Shall we base the line of demarcation on diastrophism and evaluate the fossils occurring above and below, or shall we accept the type or best-known sections and their contained faunas? On either premise—no matter which—many serious and often seemingly insurmountable difficulties arise. Tectonic disturbances do not occur simultaneously over the entire globe. The fossils that we now study lived and were entombed in the sediments of the seas which covered certain areas; at the same time, other areas were above sea level and undergoing erosion. The resulting patchwork quilt of strata and faunas must be pieced together by using all the various threads of evidence available to us.

The type Devonian is the incomplete section of strata in Devonshire, England, to which the name was hurriedly applied by English geologists in anticipation of the naming of the type section in New York State, where the best-known and most complete section is located. However, the type Kinderhook is in the Mississippi Valley, where, unfortunately, much of the excellent Devonian section present in New York is missing. Furthermore, different workers at different times have not agreed on what should be included in the Kinderhook at the type section and in adjacent areas. Some would include the black-shale formation, there called the "Grassy Creek"; others would exclude it. The unconformity below the black shale of the Mississippi Valley, considered from a diastrophic standpoint, is one of the greatest known on this continent.

Until recently our knowledge of "black-shale" stratigraphy was very patchy, and for many years little was done to correlate the numerous reports on the subject or to reconcile their differences.

In Europe, where differences of opin-

[1] Manuscript received January 15, 1948.

ion regarding the lower limit of the Carboniferous strata are as prevalent as they are in America, the conclusions reached have depended upon the class of fossils considered (Schindewolf, 1927). Coral students would place this boundary below the top of the Famennian (base of the Oberdevonstufe V), but those using cephalopods and plants place the line of separation above the base of the Tournaisian, at the base of the Unterkarbon I. Trilobites indicate that the boundary should be drawn at the base of the Tournaisian, but the brachiopod evidence indicates the base of the Strunian (K) zone. Jongmans and Gothan (1937, p. 22) selected the base of the Tournaisian Z zone as the Devonian-Carboniferous

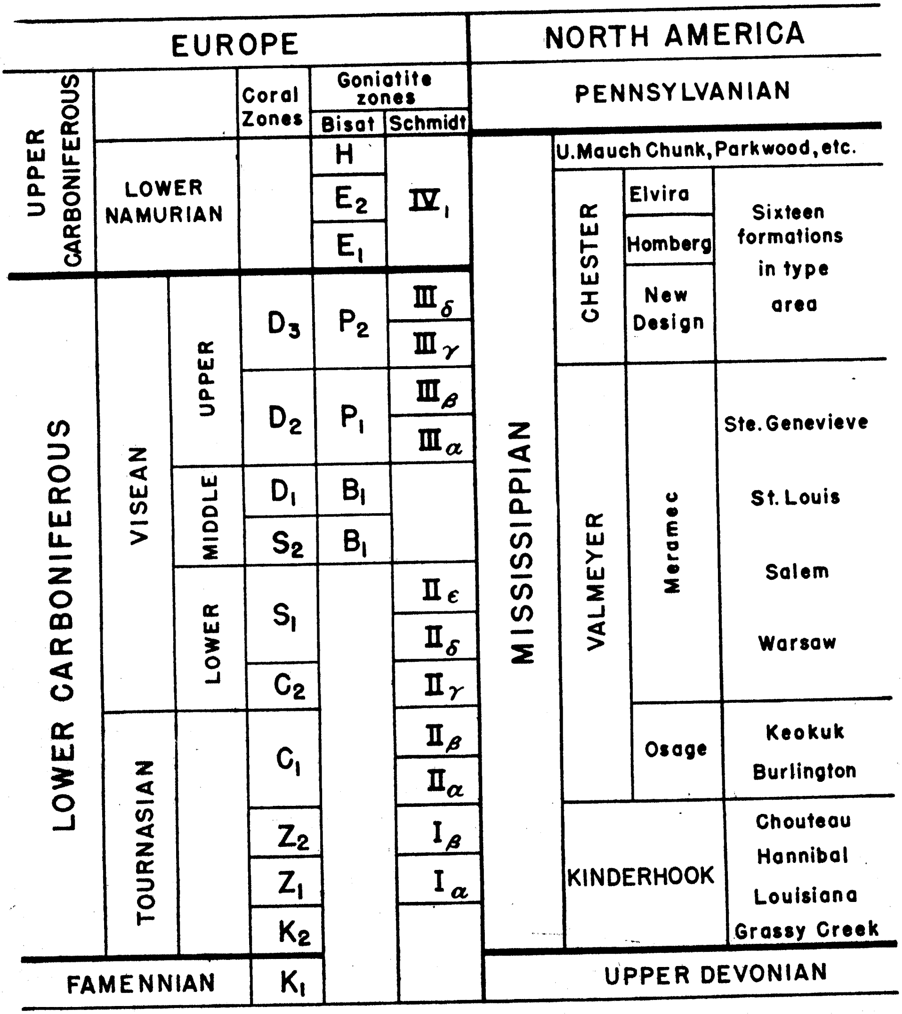



FIG. 1.—Chart showing the correlation of the American Mississippian with the Carboniferous of Europe

boundary, although the first Heerlen Congress (Jongmans, 1928) had agreed upon the base of the Strunian (K) zone (fig. 1).

In the second Heerlen Congress, a cephalopod fauna, the first appearance of which was regarded as marking the beginning of the Carboniferous, contained, among others, the genus *Protocanites*. The first appearance of this significant cephalopod genus in America is in the Louisiana limestone (Williams, 1943, p. 34).

J. Bandet (1945), in a detailed description of the stratigraphy and paleontology of the Dinantian of Tournai, shows that the calcareous shales and limestones of the *Assie de Maredosus* ($Z_{1-2}$) occur in the Lower Tournaisian. Williams (1943, p. 48) notes that individual genera in the Louisiana limestone suggest that the Louisiana is probably the time equivalent of the lower part of these beds. A photograph by Ubaghs (in Ancion, Van Leckwijck, and Ubaghs, 1943) shows (in fig. 5) thin-bedded black Namurian shales resting upon Famennien sandstones in the Liége syncline of Belgium. The systemic contact was placed below these shales.

## TYPE KINDERHOOK

F. B. Meek and A. H. Worthen (1861, p. 288) defined the type Kinderhook to include "the beds lying between the Black slate and the Burlington limestone, which have heretofore been considered the equivalents of the Chemung group of New York," which later (1866, p. 288) they said included "the Chouteau limestone, the Lithographic [Louisiana] limestone and the Vermicular sandstone and shale [Bushberg-Hannibal] of the Missouri Report, that part of the Waverly sandstone of Ohio which overlies the Black slate of that region, and the Goniatite [Rockford] limestone of Rockford, Indiana."

Workers since Meek and Worthen have advanced many correlations of the strata in outlying areas with those of the type area. Stuart Weller (1905, p. 625) correlated the Louisiana limestone with beds 2–4 in his section (1900, p. 65) at Burlington, Iowa, and with beds 2 and 3

TABLE 1

WORTHEN'S SECTION AT KINDERHOOK, ILLINOIS

| | Feet |
|---|---|
| 5. Loess capping the bluff | 20 |
| 4. Burlington limestone | 15 |
| 3. Thin-bedded, fine-grained limestone | 6 |
| 2. Thin-bedded sandstone and sandy shales | 36 |
| 1. Argillaceous and sandy shales, partly hidden | 40 |

WELLER'S SECTION AT BURLINGTON, IOWA

| | Feet |
|---|---|
| 7. Soft, buff, gritty limestone | 3– 5 |
| 6. White oölitic limestone | 2– 4 |
| 5. Fine-grained yellow sandstone | 6– 7 |
| 4. Fine-grained, compact, fragmental, gray limestone | 12–18 |
| 3. Thin band of hard, impure limestone, filled with *Chonetes;* sometimes associated with thin oölite band | ½– ¾ |
| 2. Soft, friable, argillaceous sandstone, sometimes harder and bluish in color, filled with fossils in the upper portion, the most abundant of which is *Chonopectus fischeri* | 25 |
| 1. Soft, blue argillaceous shale (exposed) | 60 |

described by Worthen (1887, p. 27) in the type area at Kinderhook, Illinois (table 1). However, R. C. Moore (1928, p. 50) believes that the Louisiana beds do not correlate with the fine-grained limestones at Kinderhook and Burlington but that beds 2 at each place correlate with the Hannibal shale of the Missouri section. L. R. Laudon (1931, p. 365) agrees with Moore's correlation of bed 2, and J. S. Williams (1943, p. 40) believes that the Louisiana limestone is older than bed 2

at Burlington. Williams' important contribution gives strong stratigraphic and faunal evidence for the Carboniferous age of the Louisiana.

A recent paper of significance in relation to Devonian-Mississippian boundary problems is that by M. A. Stainbrook (1947, p. 302) on the brachiopods from the Percha shale of New Mexico and Arizona, who says:

> The Percha shale as now recognized should not be confused with formations such as the Sly Gap and Canutillo, which contain definite Devonian faunas. The Percha fauna is more closely related to that of the overlying Mississippian than to the underlying Devonian and the Mississippian system in New Mexico seems to begin with a body of black shale, the Silver member, similar to the basal black shale of the Mississippian system in many other parts of North America.
>
> The Percha fauna is most closely related to that in the lower part of the original Ouray limestone, which has generally been considered Devonian. The upper part of the Ouray is now correlated with the Leadville, has been consistently regarded as Mississippian. If the Percha be Mississippian, the lower Ouray is Mississippian also, and the Devonian is but scantily represented in Colorado.
>
> If the Percha is Mississippian, spiriferoids of the type of *Spirifer disjunctus* are demonstrated to have persisted beyond the close of the Devonian period. This is important in connection with the age determination of some other borderline Mississippian-Devonian formations.

A few of the ostracodes of the Percha shale and others from the "Devonian-Caballero" contact in New Mexico sent me by Frank V. Stephenson (1945) are similar to those found in the Underwood member of the upper New Albany formation. A more extensive collection from the Percha and associated formations should yield important evidence for comparison with the Mississippi Valley faunas.

Wilbert H. Hass (1947*a*, p. 1190) regards the Chappel formation as Upper Devonian(?) and Mississippian because the conodonts show five zones, as follows: (1) pre-Burlington, (2) pre-Welden, (3) Chouteau, (4) Bushberg and Hannibal, and (5) Grassy Creek. The Ives breccia, immediately below the Chappel, is assigned to the Mississippian. Neither the megafaunas nor the stratigraphy justifies this zonation, which seems to demonstrate the conformability of these strata in the Mississippian (see also Barnes, Cloud, and Warren, 1947).

## NEW ALBANY SHALE

Huddle's paper on conodonts (1934) and Campbell's report on the stratigraphy of the black shales of central United States (1946) are important works and constitute a particularly complete study of the New Albany shale in its type area of southern Indiana (table 2). Although future work may alter some of Campbell's correlations with black shales in other areas some distance from the type area, his studies furnish the key to many problems that have confounded stratigraphers for many decades. Huddle's division of the New Albany into three zones, based upon conodonts, later amplified and applied to a much wider area by Campbell, proved to be the key to the stratigraphy.

Study of the relations of conodont faunas to one another and to other types of fossils, both plant and animal, now make possible the dating of at least two of the three members of the New Albany shale with reasonable certainty. The black-mud environment which began late in the Middle Devonian and persisted well into the Lower Mississippian was inhospitable to a normal marine fauna, so that, with a few exceptions, such faunas are found only at widely separated localities and horizons.

TABLE 2

CAMPBELL'S (1946) SECTION OF THE NEW ALBANY SHALE AT NEW ALBANY, INDIANA, AND VICINITY

| Unit | Feet |
|---|---|
| Rockford limestone | |
| Jacobs Chapel shale | |
| New Albany shale | |
| Henryville formation, fissile black shale, 24 species of conodonts, 16 of which are common to the lower Sanderson, D1 | 1.00 |
| Underwood formation, soft, greenish, fossiliferous shale, with a layer of phosphatic nodules at the top | 0.50 |
| Sanderson formation, D | |
| 3. Falling Run member, a layer of phosphatic concretions; *Colpocaris*, *Spathiocaris*, *Lingula melie*, *Orbiculoidea herzeri*, *Dinichthys*, *Rhadinichtys*, abundant conodonts, and 15 species of plants | 0.33 |
| 2. Black shale, *Lingula* sp., fish dermal plates and teeth, conodonts, and 8 species of plants. | 1.00 |
| 1. Black shale as in D2 with a layer of phosphatic nodules at the top; 46 species of conodonts, all common to the beds above | 9.00 |
| Blackiston formation, 60 species of conodonts, 40 species confined to the Blackiston | |
| Upper Blackiston member, C; hard, brittle, very thinly fissile black shale, with much pyrite, interbedded layers of gray shale, and calcareous concretions | |
| 4. Black shale with layers of sandstone ½–1 inch thick and flat calcareous concretions; silicified *Callixylon newberryi* | 6.50 |
| 3. Black shale, silicified *Callixylon* | 10.00 |
| 2. Black shale with large calcareous concretions | 8.00 |
| 1. Black shale with layers of gray shale and large calcareous concretions | 7.00 |
| Lower Blackiston member, B; black shale with coarse laminae, brown shale filled with *Sporangites*, and gray shale; *Barroisella campbelli* in brown shale and gray shale only | |
| 6. Brown shale crowded with *Sporangites*, thin, sheety laminae and layers of gray shale; terminated by two layers of sandstone with *Sporangites* and *Barroisella* | 5.00 |
| 5. Black-shale and gray-shale layers | 20.00 |
| 4. Black shale with prominent widely spaced joints, which extend to the layer of pyrite below | 15.00 |
| 3. *Spathiocaris* zone; gray-clay shale, which thickens to 15 feet and contains calcareous concretions at Hayden, 50 miles to the north | 0.50 |
| 2. Brownish-black to coffee-brown shale, filled with *Sporangites*, thin, sheety laminae; *Barroisella campbelli* abundant and well preserved | 1.50 |
| 1. Black shale, layer of pyrite at top, and four layers of ripple-marked sandstone 1–4 inches thick; the two middle layers contain thin lime concretions and much pyrite; *Barroisella campbelli*, *Dinichthys magnificus*, many worn fish fragments, worm borings and trails, and matted masses of unidentifiable plants | 4.50 |
| Blocher formation, A; black shale with coarse laminae; 44 species of conodonts, 39 species confined to the Blocher | |
| 3. Black shale, interbedded with layers of ripple-marked sandstone | 3.00 |
| 2. Black shale | 3.50 |
| 1. Black shale with lenses and concretions of lime at the top, lenses of sandstone at bottom; *Leiorhynchus*, *Styliolina* abundant; *Schizobolus* confined to lower 1 foot | 6.50 |
| Layer of pyrite, 1–6 inches thick, often conglomeratic, is usually found at the base of the New Albany | |
| Hamilton limestones | |

### LOWER NEW ALBANY

The lower conodont fauna of Huddle contains 44 species and occurs in the lower 8–10 feet of the black shale, named the "Blocher formation" by Campbell. Of these species, 15 are common to the Devonian of New York, 8 are found in the Rhinestreet shale, 9 in the Genundewa limestone, and 3 in the Genesee shale. Associated fossils are *Chonetes lepidus* Hall, *Leiorhynchus quadricostatus* Hall, *L. limitare* Vanuxem, *Styliolina fissurella intermittens* Hall, *Tentaculites gracilistriatus* Hall, *Lingula spatulata* Vanuxem, and *Schizobolus concentricus* Vanuxem. G. A. Cooper (1942, pp. 1773–1774) correlated these beds with the Geneseo shale and Tully limestone of the New York section on the basis of the occurrence of *Leiorhynchus*, *Schizobolus*, and *Styliolina*.

### MIDDLE NEW ALBANY

The next 75 feet of the New Albany shale in the type area is the Blackiston formation, which contains a different faunal assemblage. This is the zone of Huddle's middle New Albany conodonts, which contains 60 species, of which 3 occur in the Blocher formation, 15 in the upper New Albany shale, 10 in the Grassy Creek shale of Missouri, and 12 in the Chattanooga shale of Alabama. Nearly all the conodonts occur in the lower Blackiston formation (59 species); 19 are found in the upper part, and 18 are common to the two members. Branson and Mehl's (1933) so-called "diagnostic" Devonian genera—*Ancyrodella*, *Ancyrognathus*, *Icriodus*, *Polylophodonta*, and *Palmatolepis*—are found in this formation and, with the exception of the last, are not found in higher zones. Campbell (1946, p. 842) divided the Blackiston formation into two members on the occurrence of *Barroisella campbelli* in the lower part, where *Spathiocaris emersoni*, a new species of *Cypridinella*, and *Dinicthys magnificus* have also been identified. *Sporangites* (*Tasmanites*) are abundant in both the upper and the lower parts of the lower Blackiston formation.

Hass (1947*b*, pp. 138–139) disagrees with Campbell's correlation of the middle New Albany shale with the black-shale section in Ohio. He does not believe the Olmstead member of the Cleveland shale in northern Ohio is the correlative of the upper Blackiston formation. However, to place the Cleveland and Bedford shales at the position of an assumed hiatus in the New Albany (between the Blackiston and the Sanderson formations) disrupts a well-established correlation between the Bedford shale of Ohio, the Underwood shale of southern Indiana, and the Hamburg and Glen Park formations of the Mississippi Valley. Whether one considers this zone Devonian, Mississippian, or both in age does not invalidate this correlation.

Kindle (1900) many years ago collected *Barroisella subspatulata* and a number of goniaties from the *Spathiocaris* zone at Delphi, Indiana. Later A. K. Miller (1938) identified the goniatites as *Werneroceras wabashensis* (Kindle), *Manticoceras delphiensis* (Kindle), *M. kindlei* Miller, and *M. unduloconstrictum* Miller. G. A. Cooper (1942, p. 1738) has also found the *Barroisella-Manticoceras* association in Michigan. This well-authenticated contemporaneity seems to establish the Devonian age of *Barroisella* and thus, in turn, fixes the age of the Blackiston formation, by reason of the conodonts which occur throughout the formation. The upper Blackiston beds have yielded numerous specimens of *Callixylon newberryi* at many localities. This species is also found in the lower Ohio shale of Ohio and the Woodford forma-

tion of Oklahoma. However, the presence of *Callixylon* specifically undetermined is, in itself, no assurance of Devonian age, as will be shown later.

#### UPPER NEW ALBANY

Branson and Mehl (1938*b*, p. 129) consider the break in conodont faunas between the Grassy Creek shale and Bushberg sandstone as the greatest known in this part of the section. The same break occurs also between the Blackiston and the Sanderson, the next formation above. However, Campbell (1946, p. 852) points out that a flora, *Dinichthys*, the Underwood fauna, and 20 species of Blackiston, Blocher, and Rhinestreet conodonts also occur above this break, all of which are regarded as Devonian by some geologists. The upper New Albany shale has been divided into the Sanderson, Underwood, and Henryville members from base to top. Other fossils that have been recognized in this zone are *Lingula melie*, *Orbiculoidea herzeri*, *Cladodus springeri*, and *Rhadinichthys*, together with a new conodont assemblage containing *Synprioniodina*, *Solenognathus*, *Siphonognathus*, *Falcodus*, and *Spathodus*—all of Mississippian affinity. These genera are represented by 17 species in the upper New Albany shale, 2 in the Blackiston formation, and 1 in the Blocher formation. The Underwood and Henryville conodonts, brachiopods, and plants constitute a significant faunal zone, with the Underwood fauna and the fossiliferous concretionary zone of the Falling Run member, in the middle.

In 1939, on the basis of 63 species of conodonts, I correlated a pre-Welden–post-Woodford shale in Oklahoma with the Bushberg and Hannibal formations of the Mississippi Valley. Eighteen of these species occur in the upper New Albany shale, 5 in the Blackiston, and 2 in the Blocher formation. On the evidence of *Lingula*, *Orbiculoidea herzeri*, the *Siphonognathus* group of conodonts, and the pre-Welden and Bushberg species, the Mississippian-Devonian contact is placed at the base of the Sanderson by Campbell (1946, p. 853). However, because of the occurrence of an even greater variety of conodonts in the shale in question than is found in the pre-Welden, I now believe it is the correlative of the Jacobs Chapel shale, which lies above the upper New Albany shale and below the Rockford limestone. This shift, though slight, would correlate the upper New Albany formations with the upper Woodford beds of Oklahoma. Both contain an identical phyllocarid crustacean fauna (C. L. Cooper, 1932) in phosphatic concretions, as well as many similar species of conodonts and other classes of fossils. The concretions from the Falling Run member in Indiana have furnished the following species: *Lingula* cf. *L. cuyahoga* Hall, *Orbiculoidea herzeri* Hall, *Colpocaris bradleyi* Meek, *C. elytroides* Meek, *Tropidocaris?* sp., *Spathiocaris woodfordi* Cooper, *S. tenuicostata* Cooper, *S. striatula?* Cooper, *Angustidontus seriatus* Cooper (C. L. Cooper, 1936), *Rhadinichthys*, *Dinichthys*, *Cladodus springeri*, and other fish remains, as well as the following plants: *Kalymma lirata* Read, *Periastron perforatum* Scott and Jeffrey, *P. reticulatum* Unger, *Archeopitys eastmani?* Scott and Jeffrey, *Callixylon browni* Read, *Clepsidropsis betrandi* Read and Campbell, *C. titan?* Read, *Cladoxylon* sp., *Siderella scotti* Read, *Lepidodendron boylensis* Read, *Reimania indianensis* Read and Campbell, *Protocalamites dorfi* Read and Campbell, *Asteroxylon setchelli* Read and Campbell, *Mensoneuron simplex* Read and Campbell, *Arnoldella minuta* Read, and *Sporangites* (*Tasmanites*) *huronensis* Dawson. This flora

has been considered Mississippian by several paleobotanists (Arnold, 1947, p. 1271; Hoskins and Cross, 1947, p. 1194).

Above the nodule-bearing Falling Run member is a thin, soft, greenish-gray shale, from which Huddle (1933) reported *Chonetes yandellanus seymourensis*, *Camarotechia exima*, *Rhipidomella vanuxemi newalbanyensis*, and *Platyrachella* cf. *P. macbirdei*, assigning a Devonian age to the shale. However, G. A. Cooper (1942, p. 1774) correlated this marine zone, named the "Underwood" by Campbell, with the Hamburg beds of the Mississippi Valley and the Bedford shale of Ohio, which, together with the Glen Park limestone, contain closely related faunas.

Sixty-nine species of conodonts occur in the upper New Albany shale, of which 24 are found in the Henryville beds and 16 in the Sanderson formation. *Lingula melie* and *Orbiculoidea herzeri* occur in the Falling Run member and the Henryville, whereas 3 species of plants are found in the upper New Albany shale. The conodonts, brachiopods, and plants form a biotic unit, in the midst of which the Underwood shale is found (Campbell, 1946, p. 853).

## JACOBS CHAPEL SHALE

Above the New Albany formation lies a greenish-gray to dark-green glauconitic shale, less than a foot thick, which Campbell (1946, p. 885) named the "Jacobs Chapel shale." Faunally and stratigraphically it assumes an importance all out of proportion to its rather insignificant occurrence. In it is found again the prolific conodont fauna of a very similar zone found between the Woodford and Welden formations of south-central Oklahoma (C. L. Cooper, 1939), a locality some 700 miles away from the Jacobs Chapel locality in Indiana. As in the pre-Welden shale, most of the typical Bushberg conodonts occur in the Jacobs Chapel shale. Also found are *Meniscophyllum*, *?Chonetes*, *Rhipidomella*, *Platyceras*, minute gastropods and corals, trilobites, and ostracodes.

## ROCKFORD LIMESTONE

The youngest formation of the Kinderhook section in southern Indiana is the Rockford limestone, which immediately overlies the Jacobs Chapel shale. For the most part, the Rockford limestone is a single massive bed that may reach a thickness of 6 feet in places but is commonly 2 feet or less in thickness. Very rarely this limestone is separated into two or more beds by thin shales similar to the Jacobs Chapel shale. It is significant that these shales carry the microfauna of the Jacobs Chapel beds, which is unlike the fauna in the shales immediately overlying the Rockford limestone.

## STRATIGRAPHIC SUMMARY

In summary, the New Albany shale, together with other pre-Osage formations, comprises eight faunal and stratigraphic zones in Clark County, Indiana, and vicinity. These are, from the base: (1) the Blocher formation (Devonian), (2) the Blackiston formation, presumed to be Devonian because of the association of *Barroisella campbelli* with *Manticoceras* in the Delphi shale of northern Indiana and with *Tornoceras* in the Antrim shale of Michigan, (3) the Sanderson formation (Mississippian), which contains 43 of the 69 upper New Albany conodonts, with (4) the Falling Run member at the top containing phosphatic concretions with their characteristic crustaceans and plants, (5) the Underwood shale with its Hamburg-Bedford megafauna and an exotic undescribed microfauna, (6) the Henryville black

shale containing 24 species of the upper New Albany conodonts, (7) the Jacobs Chapel shale with its distinctive microfauna, and (8) the Rockford limestone, correlative with the Chouteau limestone of the Mississippi Valley section. Of these eight zones, 3–8 are with confidence assigned to the Kinderhook series. There is still some question about the age of zone 2 and its correlative, the Grassy Creek shale of Missouri. Zone 1 is thought to be definitely Devonian.

## KINDERHOOK MICROFOSSILS

### CONODONTS

All students of Paleozoic microfaunas are well acquainted with the conodont work of Branson and Mehl and their associates and students at the University of Missouri. Their first comprehensive work appeared in volume 8 of the *University of Missouri Studies* in 1933–34, wherein were described, among others, the conodont faunas of the Grassy Creek, Bushberg, and Hannibal formations. Later several papers appeared in the *Denison University Bulletin: Journal of the Scientific Laboratories* (1940*a;* 1940*b;* 1940*c;* 1940*d*), and the *Journal of Paleontology* (1938*a;* 1941), of which "New and little known Carboniferous conodont genera" (1941) is of special significance to this discussion. All of this work is summed up in Branson's *Geology of Missouri* (1944), which includes plates and fossil lists from unpublished theses written by graduate students at the University of Missouri, together with lists from a "phantom paper" by Branson and Mehl.[2]

The important work of Huddle (1934) on the New Albany shale has been mentioned, and the results of my own work on these fossils were given in several papers in the *Journal of Paleontology* (1931*a*, 1931*b*, 1935, 1939, 1945, 1947; Cooper and Sloss, 1943). Wilbert H. Hass currently working on conodonts from the Mississippian and Devonian formations of Texas, is the author of several recent contributions (1941, 1943; Knechtel and Hass, 1938; Hass and Lindberg, 1946). Except for papers on the Carboniferous published over half a century ago by G. J. Hinde (1879, 1900) and associates in Great Britain, these constitute the conodont literature of the Kinderhook.

There are, however, a number of undescribed faunas in my collections, including the prolific Jacobs Chapel fauna, the fauna from the shale breaks of the Rockford limestone of Indiana, that from the Kinderhook above the Mountain Glen shale and below the Springville formation of Union County, Illinois, and that from the phosphatic concretions of the Falling Run member in Indiana and Kentucky.

S. P. Ellison, Jr. (1946), has listed the stratigraphic ranges of most of the well-known conodont genera, and his chart shows all conodonts from the Grassy Creek and correlative formations below the Devonian-Mississippian line. However, listed in the Kinderhook are many of the common Bushberg, Hannibal, pre-Welden, and Jacobs Chapel genera, such as *Pinacognathus*, *Subbryantodus*, *Pseudopolygnathus*, *Siphonodella*, and *Solenodella*. The genera *Doliognathus*, *Scaliognathus*, and *Bactrognathus*, also shown in the Kinderhook, are indicative of a break in conodont faunas above the Rockford limestone that is equal or greater in importance than that claimed for the hiatus between the Grassy Creek fauna and those above it. The character-

[2] This paper, long delayed in publication (it was to have been published in 1943), will appear in the *Journal of Paleontology*. It lists several species, genera, and families from this report and from Shimer and Shrock's *Index Fossils of North America*.

istic Kinderhook species of the genera *Siphonodella*, *Solenodella*, *Macropolygnathus*, and nearly all the large number of species of *Polygnathus* are absent from the Rockford. Branson and Mehl (1941) described the new genera, *Doliognathus*, *Bactrognathus*, *Staurognathus*, and *Solenognathus*, from the Pierson of Missouri and the "Sycamore" of Oklahoma. Six species of the first two genera, together with many other new forms, are found in the shale overlying the Rockford and thus prove the Osage age of the Pierson and the limy shales above the Weldon of Oklahoma. Marvin Weller[3] believes the Welden limestone to be of Kinderhook age because similar species of trilobites occur in this formation and the Chouteau limestone of the Mississippi Valley.

The use of conodonts as index fossils has been criticized on the grounds that these fossils have not yet been properly evaluated and that their stratigraphic ranges are not sufficiently known to make them reliable markers. Many irregularities of occurrence or abnormal stratigraphic associations, such as "stratigraphic admixtures," "stratigraphic leaks," or "phantom formations," have been noted by some workers (Branson and Mehl, 1940*c*) in these fossils. However, as Guy Campbell (1946, p. 855) has pointed out, in the case of the New Albany shale, "failure to interpret the New Albany stratigraphy is due to incomplete and incorrect knowledge." It seems probable that in other instances seemingly abnormal conodont occurrences might be explained in the same manner or that a longer stratigraphic range should be recognized for some genera and species. It is indeed difficult to understand how "Welden conodonts were incorporated in the sub-Welden clay."

[3] Personal communication.

## OSTRACODES

Scarcity of published papers, both in Europe and in America, makes this class of fossils less useful in the correlation of Kinderhook beds than they might be, in spite of a complete and presumably ostracode-bearing section on both continents. In 1939, E. Kummerow (p. 74), in comparing Lower Carboniferous ostracodes with those from adjacent formations stated:

> A comparison of the ostracode fauna of the Lower Carboniferous with that of the Upper Devonian becomes difficult owing to the imperfect knowledge of the latter. Only a part (Aparchitidae, Primitiidae, Zygobolbidae, Beyrichiidae, Kloedenellidae, Entomidae) have been worked in recent time (Matern, 1929). . . . Almost every type of the now well-known *Entomis*, except the genus *Richterina*, are found also in the Lower Carboniferous. Likewise near-related forms of *Entoprimitia* and graphiodactylids (*Primitia* and *Primitiella* of Matern) occur also in both formations. . . . Through the discovery of numerous species of *Entomis* in the shales of Dasberg near Neviges (Rheinland) and in the siliceous beds of the Huttales near Claustal (Hartz) was also the statement of Matern which disproved that *Entomis* species died away at the conclusion of the Dasberg stage and the *Richterina* species at the end of the Hangenberg stage. In comparison, a change in the ostracode fauna took place after the expiration of the Hangenberg stage.

Therefore, the important change in ostracode faunas in Europe occurred at the end of the Hangenberg stage, which is in the Tournaisian. In America the genus *Entomis* is known only from the Olentangy shale of Ohio (Stewart and Hendrix, 1945) and the Alfred shale member of the "Chemung" (Canadaway) of New York.

No Upper Devonian ostracode faunas—in fact, none younger than Hamilton (Warthin, 1934; Van Pelt, 1933; Coryell and Malkin, 1936; Stewart, 1936)—have been described from North America. Only two papers on Kinderhook ostra-

codes, both by P. S. Morey (1935,[4] 1936) have been published in this country; one describes a fauna from the basal Mississippian sandstone (Sylamore) and the other a fauna from the Chouteau limestone, both from outcrops in Missouri. A third paper, by R. S. Bassler (1932, p. 236) is on a fauna which was thought to be of Kinderhook age at the time that the paper was published but is now known to be of lower Osage age (Wilson and Spain, 1936). In my own collections are undescribed faunas from Texas, Oklahoma, Missouri, and Indiana.

These reports show the presence of several holdover Devonian-like forms, such as *Dizygopleura mehli* Morey, two species of the new genus *Plagiophrenodes*, and two species of small, flat, deeply punctuate *Amphissites*, without nodes or ridges, the latter somewhat similar to Hamilton forms. Carry-over genera from the Devonian also include new species of *Bairdia*, *Healdia*, *Graphiodactylis*, the latter represented by several species in the pre-Welden beds of Oklahoma, the Chappel formation of Texas, the pre-Springville (Darty) limestone of Illinois, and the Jacobs Chapel shale of Indiana.

The exotic fauna from the Hamburg-Bedford faunal zone in the Underwood shale of Indiana has, as far as known to me, no counterpart in Devonian or Carboniferous faunas of North America or Europe. It contains a terminal-spined form quite similar to *Ropolonellus*, although apparently it is not *R. papillatus* Van Pelt, from the Hamilton of Michigan and Ontario. The fauna also contains species of *Graphiodactylis* and *Kirkbyella*. Two forms similar to *Waylandella* and *Cornigella* give a strong Carboniferous aspect to this fauna.

[4] The horizon considered in this article (basal Mississippian) has been correlated with the Bushberg sandstone.

There are no Osage ostracodes known to me other than those from the Ridgetop shale of Tennessee, which contains the Devonian-like *Aechmina longicornis* (Ulrich and Bassler) and species of the ubiquitous *Tetrasacculus*, which first makes its appearance in the Hamilton beds of Ohio and Ontario, next in the Lower Louisiana limestone of Missouri, then in the Ridgetop formation, and finally high in the Mississippian in the Golconda and Glen Dean formations of the Chester series (C. L. Cooper, 1941) of the Mississippi Valley. The next higher ostracodes above the Osage are those described by H. L. Geis (1932) from the Salem (Meramec) limestone of Indiana.

### FORAMINIFERA

Although Foraminifera are known to occur in many Mississippian formations, the only described forms are *Endothyra baileyi* from the Salem and St. Louis beds and a few genera from the Kinkaid (C. L. Cooper, 1947) (Chester) beds of Illinois which include species of *Millerella*, *Endothyra*, *Hyperammina*, *Palaeotextularia*, and *Trepeilopsis*.

Foraminifera have been observed in the Louisiana, Jacobs Chapel, Chouteau, and Rockford formations; a species of *Endothyra*, similar to *E. baileyi*, was collected from the pre-Welden shale. These faunas are all undescribed. In the Louisiana limestone are two species of *Ammodiscus*, while the Jacobs Chapel shale and Rockford beds contain many species of *Trepeilopsis*, *Glomospira*(?), and one species of *Thurammina*. Many identical forms also occur in the Chouteau limestone.

The Louisiana limestone is reported (Williams, 1943, p. 55) to contain forms that resemble *Hyperamminoides*, *Lituotuba*, and *Ammodiscus*.

### MICROCRINOIDS

*Allagecrinus americanus* Rowley is common in the Louisiana limestone of Illinois and Missouri. Similar forms, but ones which may prove to belong to different genera and species, have been observed in the pre-Welden strata of Oklahoma, the Jacobs Chapel shale of Indiana, and the Chappel formation of Texas. Dissociated plates, probably of *Poteriocrinus* or *Platycrinus*, occur in the same zones.

### OTHER MICROFOSSILS

Spores are abundant in the black shales of the New Albany type (Schopf, Wilson, and Bentall, 1944). *Sporangites* (= *Tasmanites*) *huronense* is the common and only described species, but further study will probably reveal others. This type of spore is known to occur in rocks as old as Ordovician.

Radiolaria have been reported (Henbest, 1936; Aberdeen, 1940) from the Caballos and Arkansas novaculites of Texas and Oklahoma. Designated as Devonian(?) by the United States Geological Survey, these formations are thought to be correlative with some part of the New Albany formation, probably the Blocher or the Blackiston, or both.

Holothurian plates, though undescribed in the Kinderhook, are represented by the numerous round, lacelike disks (*Ancistrum?*) found in many zones. Plates and other structural parts are known from higher Mississippian beds (Croneis and McCormack, 1932).

A few echinoid spines of an undetermined species have been noted in the Jacobs Chapel shale and correlative formations.

Annelida are represented in the Louisiana limestone (Williams, 1943, pp. 61–63) and Jacobs Chapel shale; *Conularia marionensis* Swallow, *Spirorbis kinderhookensis* Gurley, and *Cornulites carbonarius* Gurley are common in the Louisiana limestone. Scolecodonts are rare, but a few have been observed in the Underwood member of the upper New Albany shale.

Many nepionic forms show up in the microsamples, such as brachiopods, probably *Schuchertella louisianensis* Williams, gastropods, and ammonites, probably *Protocanites louisianensis* (Rowley).

## REFERENCES CITED


ABERDEEN, E. J. (1940) Radiolarian fauna of the Caballos formation, Marathon Basin, Texas: Jour. Paleontology, vol. 14, pp. 127–139.

ANCION, CH.; LECKWIJCK, W. VAN; and UBAGHS, G. (1943) A propos de la bordure méridionale du synclinal de Liége, à l'aval de Liége: la ride famennienne de Booze–Le Val Dieu, à la limite septentrionale du plateau de Herve: Soc. géol. Belgique Annales, vol. 66, pp. 299–335.

ARNOLD, C. A. (1947) The Mississippian flora (abstr.): Geol. Soc. America Bull. 58, p. 1271.

BANDET, J. (1945) Paléontologie stratigraphique du calcaire Dinantien du Tournaisis: Soc. géol. France Compte rendu, ser. 5, vol. 15, pp. 633–638.

BARNES, V. E.; CLOUD, P. E.; and WARREN, L. E. (1947). Devonian rocks of central Texas: Geol. Soc. America Bull. 58, pp. 125–140.

BASSLER, R. S. (1932) Stratigraphy of the Central Basin of Tennessee: Tennessee Geol. Survey Bull. 38.

BRANSON, E. B. (1944) The geology of Missouri: Univ. Missouri Studies, vol. 19, no. 3.

———, and MEHL, M. G. (1933) Conodonts from the Grassy Creek shale of Missouri: Univ. Missouri Studies, vol. 8, pp. 171–259.

——— — ——— (1938a) The conodont genus *Icriodus* and its stratigraphic distribution: Jour. Paleontology, vol. 12, pp. 156–166.

——— — ——— (1938b) Stratigraphy and paleontology of the Lower Mississippian of Missouri: Univ. Missouri Studies, vol. 13.

——— — ——— (1940a) Caney conodonts of Upper Mississippian age: Denison Univ., Sci. Lab., Bull. 35, pp. 167–178.

——— — ——— (1940b) Conodonts from the Keokuk formation: *ibid.*, pp. 179–188.

——— — ——— (1940c) The recognition and interpretation of mixed conodont faunas: *ibid.*, pp. 195–209.

——— — ——— (1940d) A record of typical American conodont genera in various parts of Europe: *ibid.*, pp. 189–194.

——— — ——— (1941) New and little-known Carboniferous conodont genera: Jour. Paleontology, vol. 15, pp. 97–106.

CAMPBELL, GUY (1946) New Albany shale: Geol. Soc. America Bull. 57, pp. 827–908.

COOPER, C. L. (1931a) Conodonts from the Arkansas novaculite, Woodford formation, Ohio shale and Sunbury shale: Jour. Paleontology, vol. 5, pp. 143–151.

——— (1931b) New conodonts from the Woodford formation of Oklahoma: *ibid.*, pp. 230–243.

——— (1932) A crustacean fauna from the Woodford formation of Oklahoma: *ibid.*, vol. 6, pp. 346–352.

——— (1935) Conodonts from the upper and middle Arkansas novaculite, Mississippian, at Caddo Gap, Arkansas: *ibid.*, vol. 9, pp. 307–315.

——— (1936) Actinopterygian jaws from the Mississippian black shales of the Mississippi Valley: *ibid.*, vol. 10, pp. 92–94.

——— (1939) Conodonts from a Bushberg-Hannibal horizon in Oklahoma: *ibid.*, vol. 13, pp. 379–422.

——— (1941) Chester ostracodes in Illinois: Illinois Geol. Survey Rept. Inv. 77.

——— (1945) Devonian conodonts from northwestern Montana: Jour. Paleontology, vol. 19, pp. 612–615.

——— (1947) Upper Kinkaid (Mississippian) microfauna from Johnson County, Illinois: *ibid.*, vol. 21, pp. 81–94.

——— and SLOSS, L. L. (1943) Conodont fauna and distribution of Lower Mississippian black shale in Montana and Alberta: *ibid.*, vol. 17, pp. 168–176.

COOPER, G. A. *et al.* (1942) Correlation of the Devonian sedimentary formations of North America: Geol. Soc. America Bull. 53, pp. 1729–1794.

CORYELL, H. N., and MALKIN, D. S. (1936) Some Hamilton ostracodes from Arkona, Ontario: Am. Mus. Novitates 891, pp. 1–20.

CRONEIS, CAREY, and MCCORMACK, JOHN (1932) Fossil Halothuroidea: Jour. Paleontology, vol. 6, pp. 111–148.

ELLISON, S. P., JR. (1946) Conodonts as Paleozoic guide fossils: Am. Assoc. Petroleum Geologists Bull. 30, pp. 93–110.

GEIS, H. L. (1932) Some ostracodes from the Salem limestone, Mississippian, of Indiana: Jour. Paleontology, vol. 6, pp. 149–188.

HASS, W. H. (1941) Morphology of conodonts: Jour. Paleontology, vol. 15, pp. 71–81.

——— (1943) Corrections to the Kinderhook conodont fauna, Little Rocky Mountains, Montana: *ibid.*, vol. 17, pp. 307–308.

——— (1947a) Conodont faunas, Central Mineral Region, Texas (abstr.): Geol. Soc. America Bull. 58, pp. 1190.

——— (1947b) Conodont zones in Upper Devonian and Lower Mississippian formations of Ohio: Jour. Paleontology, vol. 21, pp. 131–141.

——— and LINDBERG, M. L. (1946) Orientation of the crystal units of conodonts: Jour. Paleontology, vol. 20, pp. 501–504.

HENBEST, L. G. (1936) Radiolaria in the Arkansas novaculite, Caballos novaculite, and Big Fork chert: Jour. Paleontology, vol. 10, pp. 76–78.

HINDE, G. J. (1879) On conodonts from the Chazy and Cincinnati group of the Cambro-Silurian, and from the Hamilton and Genesee shale divisions of the Devonian, in Canada and the United States: Geol. Soc. London Quart. Jour., vol. 35, pp. 351–369.

——— (1900) Scotch Carboniferous conodonts: Nat. History Soc. Glasgow Trans., vol. 5, pp. 338–346.

HOSKINS, J. H. and CROSS, A. T. (1947) Survey of certain Devonian-Mississippian transition floras. Part I, Geological considerations; part II, Paleobotanical considerations (abstr.): Geol. Soc. America Bull. 58, p. 1194.

HUDDLE, J. W. (1933) Marine fossils from the top of the New Albany shale of Indiana: Am. Jour. Sci., 5th ser., vol. 25, pp. 303–314.

——— (1934) Conodonts from the New Albany shale of Indiana: Am. Paleontology Bull. 21.

JONGMANS, W. J. (ed.) (1928) Congrès pour l'avancement des études de stratigraphie Carbonifère, Heerlen, 1927, Liége, Imprimerie Vaillant-Carmanne S. A.

——— and GOTHAN, W. (1937) Betrachtungen über die Ergebnisse des zweiten Kongresses für Karbon-stratigraphie: Deuxième Congrès pour l'avancement des études de stratigraphie Carbonifère, Heerlen, 1935, Maestricht, Imprimerie Gebrs. van Aelst.

KINDLE, E. M. (1900) Devonian fossils and stratigraphy of Indiana: Indiana Dept. Geology and Nat. Resources, 25th Ann. Rept., pp. 529–758.

KNECHTEL, M. M., and HASS, W. H. (1938) Kinderhook conodonts from Little Rocky Mountains, northern Montana: Jour. Paleontology, vol. 12, pp. 518–520.

KUMMEROW, E. (1939) Die Ostracoden und Phyllopoden des deutschen Untercarbons: Preuss. geol. Landesanstalt Abh. 194.

LAUDON, L. R. (1931) The stratigraphy of the Kinderhook series of Iowa: Iowa Geol. Survey, vol. 35.

MATERN, HANS (1929) Die Ostrocoden des Oberdevons: Preuss. geol. Landesanstalt Abh. 118, pp. 1–99.

MEEK, F. B., and WORTHEN, A. H. (1861) On the age of the Goniatite limestone: Am. Jour. Sci., ser. 2, vol. 32, pp. 167–176.

MEEK, F. B., and WORTHEN, A. H. (1866) The sub-Carboniferous limestone series (of Illinois): Illinois Geol. Survey, vol. 1.

MILLER, A. K. (1938) Devonian ammonoids of America: Geol. Soc. America Special Paper 14.

MOORE, R. C. (1928) Early Mississippian formations in Missouri: Missouri Bur. Geology and Mines, ser. 2, vol. 21.

MOREY, P. S. (1935) Ostrocoda from the basal Mississippian sandstone in central Missouri: Jour. Paleontology, vol. 9, pp. 316–326.

——— (1936) Ostracoda from the Chouteau formation of Missouri: *ibid.*, vol. 10, pp. 114–122.

SCHINDEWOLF, O. H. (1928) Die Liegendgränze des Karbons im Lichte biostratigraphisher Kritik: Congrès pour l'avancement des études de stratigraphie Carbonifère, Heerlen 1927, Liége, 1928, pp. 651–661.

SCHOPF, J. M.; WILSON, L. R.; and BENTALL, RAY (1944) An annotated synopsis of Paleozoic fossil spores and the definition of generic groups: Illinois Geol. Survey Rept. Inv. 91.

STAINBROOK, M. A. (1947) Brachiopoda of the Percha shale of New Mexico and Arizona: Jour. Paleontology, vol. 21, pp. 297–329.

STEPHENSON, F. V. (1945) Devonian of New Mexico: Jour. Geology, vol. 53, pp. 217–245.

STEWART, G. A. (1936) Ostrocodes of the Silica shale, Middle Devonian, of Ohio: Jour. Paleontology, vol. 10, pp. 739–763.

——— and HENDRIX, W. E. (1945) Ostracoda of the Olentangy shale, Franklin and Delaware counties, Ohio: Jour. Paleontology, vol. 19, pp. 96–115.

VAN PELT, H. L. (1933) Some ostracodes from the Bell shale, Middle Devonian, of Michigan: Jour. Paleontology, vol. 7, pp. 325–342.

WARTHIN, A. S. (1934) Common ostracodes of the Traverse group: Univ. Michigan Mus. Paleontology Contr., vol. 4, pp. 205–226.

WELLER, STUART (1900) The succession of fossil faunas in the Kinderhook beds at Burlington, Iowa: Iowa Geol. Survey, vol. 10.

——— (1905) The northern and southern Kinderhook faunas: Jour. Geology, vol. 13, pp. 617–634.

WILLIAMS, J. S. (1943) Stratigraphy and fauna of the Louisiana limestone of Missouri: U.S. Geol. Survey Prof. Paper 203.

WILSON, C. W., Jr., and SPAIN, E. L., JR. (1936) Age of the Mississippian Ridgetop shale of central Tennessee: Am. Assoc. Petroleum Geologists Bull. 20, pp. 805–809.

WORTHEN, A. H. (1887) Geology of Pike County, Illinois: Illinois Geol. Survey, vol. 4.

# THE MISSISSIPPIAN FLORA[1]

CHESTER A. ARNOLD
University of Michigan

ABSTRACT

The Mississippian flora is one of the least investigated fossil floras of North America. David White recognized two phases—a lower, or Pocono, phase and an upper, or Chester, phase. The Pocono flora is characterized by *Triphyllopteris* and lycopods of the *Lepidodendropsis* type. The Chester flora is set off by the occurrence of *Cardiopteris polymorpha*, lycopods resembling the Old World *Lepidodendron volkmannianum*, and *Asterocalamites*. Both of these floras are distinct from the late Devonian below and the early Pennsylvanian above. The Reeds Spring formation of Missouri and the Wedington sandstone member of the Fayetteville shale contain the only Mississippian land floras yet described from the Mid-Continent region. The plants of the latter support the middle or early late Chester age indicated by the inverterbates. The flora of about forty species in the upper New Albany shale, described as Devonian, contains at least as many Mississippian elements. Most of the species found there, however, are restricted to this zone and consequently are of limited value in determining the age of the plant beds.

Although the existence of plants in the Mississippian has been known for more than a century, the flora of the rocks of this epoch has received relatively little attention. Not only is the flora very imperfectly known, but there is at present no one who is recognized as an authority on the Mississippian flora or who shows any immediate inclination to make it a subject of systematic investigation. Compared with the Pennsylvanian flora, the Mississippian flora is a small one. Localities where well-preserved Mississippian plants can be collected are few. Although some species occur in great abundance, the number of species that can be collected at any one place is small.

With the passing of the late Dr. David White, the largest and most comprehensive store of information on Mississippian plants ever possessed by one individual was irretrievably lost. During his lifetime Dr. White collected widely in the Mississippian, but his publications on the flora are few. His collections, however, are preserved in the United States National Museum, and it is earnestly hoped that some day they will be studied and described.

[1] Manuscript received January 3, 1948.

Dr. White set forth the general features of the Mississippian flora in the Appalachian trough in a paper published by the West Virginia Geological Survey in 1926. Except for changes in the names of some of the plants, the situation as White outlined it then remains essentially unaltered today. The few remarks I shall make are to a large extent restatements of what has been said before. White recognized a lower, or Pocono, phase in the Mississippian flora and an upper, or Chester, phase. These two are as distinct from each other as they are from the late Devonian below and the lowermost Pennsylvanian above.

The lower, or Pocono, flora, in addition to possessing obvious characteristics of its own, is clearly a derivative of the flora of the late Devonian. It is characterized throughout its entire lateral extent from eastern Pennsylvania to Tennessee by *Triphyllopteris*, a plant of fernlike aspects but of unknown affinities. The several species that have been named are difficult to distinguish. The Upper Devonian, on the other hand, is marked by *Archaeopteris*, which is used as a guide fossil in both North America and Europe. At almost any plant locality

in the Pocono and Price formations one may find fragments of the round-pinnuled plant resembling *Aneimites acadica*, described by Dawson from Horton Bluff, and which has been found across the Atlantic in the Cementstone group of the Calciferous sandstone of Great Britain. Some paleobotanists, including Robert Kidston, have regarded *Aneimites* and *Triphyllopteris* as congeneric and have advocated the use of the former name for both on grounds of priority. Whether they are distinct genera, I am not prepared to say, although it is evident even from casual examination that they intergrade. Other leaf types, bearing such names as *Rhodea*, *Cardiopteridium*, *Neurocardiopteris*, and *Sphenopteris*, are present in the Pocono flora. These will not be further commented upon.

The Lower Mississippian flora contains distinctive lycopods which are evidently derivatives of the Devonian *Protolepidodendron* and *Archaeosigillaria* and with which they are easily confused when not well preserved. The so-called "*Lepidodendron*" *corrugatum* described by Dawson from Horton Bluff is the same type of lycopod as "*L.*" *scobiniforme*, discovered by Meek in the Pocono. These lycopods are not true *Lepidodendra* because the cushions lack the characteristic oblique arrangement and the distinctive leaf-scar markings. They have recently been referred to the genus *Lepidodendropsis*, described by Lutz from the Culm of Bohemia. Imprints of true lepidodendrids and sigillarians have not been found in the early Mississippian; *Cordaites* has not been found in the Pocono or Price; and calamitean remains are fragmentary and scarce.

During a tour of the United States in 1933, Jongmans collected plants in the Pocono and Price formations. In collaboration with Gothan and Darrah, Jongmans (1937) described 11 new species and made a few generic transfers. Unfortunately for any American paleobotanist who may undertake a restudy of this flora, the holotypes of Jongmans' species are in the Dutch Geological Bureau at Heerlen. Jongmans, however, collected mostly at well-known accessible localities, so that it may be possible to designate topotypes of most of his species, although poor figures of some render their future identification virtually impossible.

In the Appalachian trough we know less of the Upper Mississippian flora than we do of the Lower, but the two stand out as distinct from each other. Most characteristic of the upper flora is a species of *Cardiopteris* which was assigned by White to *C. polymorpha*, well known in the late Lower Carboniferous of the Old World. Lycopods resembling the Old World *Lepidodendron volkmannianum* make an appearance, as does also *Asterocalamites*, which occurs in the late Lower Carboniferous elsewhere.

In the late Upper Mississippian a few forerunners of Pennsylvanian species come in, but only in the very latest of the deposits do species occur which render paleobotanical distinctions between the terrains difficult.

The most important contribution to the late Mississippian flora is White's account, published posthumously, of the flora of the Wedington sandstone member of the Fayetteville shale of Arkansas (1937*a*). The flora as described contains thirty-five species of ferns, seed-ferns, lycopods, calamites, and sphenophylls. Most of the forms are preserved as fragments, but they are sufficiently distinctive to support the Middle or early Upper Chester age indicated by the invertebrates. The flora is also similar in many

ways to standard Lower Carboniferous floras of Europe. Its Mississippian aspect is modified to some aspect by early Pennsylvanian and Namurian elements, but it contains no forms in common with the Middle Pottsville (Morrow) flora that occurs in the same region.

The Stanley shale and Jackfork sandstone of western Arkansas and southeastern Oklahoma were originally regarded by White as very late Mississippian. However, a careful analysis (completed in 1932) of a very fragmentary flora from these beds convinced Dr. White that, although they are older than the coal-bearing shale member of the Morrow, they are younger than any known Mississippian formations in eastern North America (1937*b*). Dr. White concluded that the Stanley and Jackfork are equivalent to the lower part of the Morrow group and hence are early Pennsylvanian.

Very few direct comparisons have been drawn between the Mississippian floras of the United States and floras of similar age in Europe, but it is generally assumed that they closely approximate the Lower Carboniferous floras of Great Britain and those of the Dinantian of the Continent. Jongmans (in Jongmans and Gothan, 1933) calls attention to the common occurrence of *Lepidodendropsis* in the Pocono and Price and the early Viséan of Bohemia. White (1913) has shown that the flora of the Pocono and Price is equivalent to that of Horton Bluff in Nova Scotia, which, in turn, ties up with the Calciferous sandstone flora of Scotland. Bell (1929) adopts White's assumptions and, in addition, correlates the Windsor flora with those of the Chester and the Middle and Upper Viséan.

Any survey of the Mississippian flora should include mention of the petrified woody stems described by the late Dr. J. E. Cribbs from the Reeds Spring formation of southwestern Missouri. Dr. Cribbs gave a detailed account of four types, one a species of *Cordaites* but the others representing genera previously unknown (Cribbs, 1935, 1938, 1939, 1940).

In connection with the foregoing account of the Mississipian flora, it seems appropriate to add a few remarks on the New Albany shale. Ever since 1850 the shale has been known to contain large petrified tree trunks. Recently, Charles B. Read and Guy Campbell (1939) have described a flora from the shale totaling forty-three species.

The large tree trunks in the New Albany shale belong, as far as we know, to one species, *Callixylon newberryi*. According to Campbell's recent work (1946), these trunks occur in a 16½-foot zone at the top of the Upper Blackiston member of the Blackiston formation. Since Campbell (1946) draws the Mississippian-Devonian boundary at the top of the Upper Blackiston member, *C. newberryi* is in the Devonian phase of the shale. However, fragments of *Callixylon* wood which appear to be specifically different from *C. newberryi* occur in the Sanderson formation which overlies the Blackiston. Therefore, if Campbell is correct in drawing the boundary where he does, *Callixylon* ranges beyond the Upper Devonian, where it principally occurs, into the Mississippian. This casual wandering of the genus beyond its alloted bounds, however, does not seriously impair its significance as a guide fossil, because it is still essentially a Devonian genus.

Most of the New Albany shale plants described by Read and Campbell come from the Falling Run member of the Sanderson, which, according to Camp-

bell's latest paper (1946), is Mississippian. This flora is a most remarkable one, not only for the variety of plant types present but for the fact that the plants are histologically preserved. In this respect it presents a marked contrast to others of comparable age in North America which are made up principally of compressions and impressions. This type of preservation (petrifaction) renders the flora one of exceptional interest to the investigator concerned with the anatomy of ancient plants. However, from the stratigraphical standpoint, such a flora presents special problems, and it is of more limited value than is a flora preserved in the more usual form. To be used for stratigraphic purposes, it should be compared with other floras preserved in a similar manner, and in North America no such floras of comparable age exist. As a consequence of the type of preservation of the plant remains, the New Albany shale stands quite alone as a landmark in the panorama of ancient plant life, just as the whole black-shale series is a sedimentary unit that is different in many ways from other sedimentary units of comparable age.

On the basis of six genera, Read (1936) and Read and Campbell in a joint paper (1939) assign the Sanderson (Upper New Albany) flora to the Devonian. These genera are *Asteroxylon*, *Polyxylon*, *Pietzschia*, *Protolepidodendron*, *Reimannia*, and *Callixylon*.

Before discussing the suitability of these six genera as age indicators in this particular instance, I wish to interject a general statement embodying a principle. When fossils are used in stratigraphic correlation, only those are suitable which have been previously observed to be restricted to, or principally restricted to, certain zones or happen to be confined within well-defined limits. At best, an element of relativity clings to the application of the concept of a guide fossil because detailed studies of newly discovered faunas or floras almost invariably result in an extension of the previously known range of some of the species. It may be said that a guide fossil remains a guide fossil only until it has been discovered at some place where it is not supposed to exist. With these thoughts in mind, we shall refer again to the six genera which Read and Campbell claim are sufficient to render the Upper New Albany flora "distinctly Upper Devonian."

If *Callixylon* were the only plant in the Sanderson formation and there were no contradictory evidence, we could assign the Sanderson to the Upper Devonian with complete assurance. However, Read and Campbell name about forty plants from this bed or its equivalent in southern Indiana and central Kentucky, from which they select five, in addition to *Callixylon*, which they claim are sufficient to place it in the Devonian. Among these, *Asteroxylon* is named, a genus previously reported twice, once from Scotland and again from western Germany. Both these occurrences are in the Middle Devonian, and they furnish all the data that we have concerning the stratigraphic range of the genus. The plant identified as *Asteroxylon* is illustrated by one figure of the transverse section of a stem containing a five-lobed wood strand which resembles that of *Asteroxylon* but which, in the absence of leaves and other essential characters, cannot be positively identified. There are many other ancient plants that have furrowed wood strands which might, on the basis of a cross section alone, be confused with *Asteroxylon*. *Pietzschia* has been reported once from the Upper Devonian of central Europe, but we probably know less of its range than we do of *Asteroxylon*. *Protolepido-*

*dendron* is a Middle and Lower Devonian plant known mainly from compressions of stems bearing numerous short bifurcated leaves. There are too many other lycopods similar to *Protolepidodendron* of whose structure we know nothing to use any of them as guide fossils. *Lepidodendropsis*, for example, is a Mississippian lycopod probably derived from *Protolepidodendron*, and we should take into consideration the likelihood of the two having similar structure. *Reimannia* has been known (until reported from the New Albany shale) only from the Hamilton of New York, and our entire knowledge of this genus was based upon one specimen. It was interpreted as the prototype of a well-known group of Carboniferous ferns, and consequently it is not surprising to find the same or a very similar plant in the New Albany shale. It is readily admitted that all the plants discussed above are suggestive of the genera to which they have been referred, but the published figures are the main basis for judgment. Either more material must be examined, or that which has been examined must be more completely described, before a final verdict can be given and we can be absolutely certain of their generic identity or of their value as age determiners.

Another fact that calls for caution in using the plants discussed above in dating the New Albany shale is that the whole flora is peculiar to this series. Of the forty-three plants listed by Read and Campbell, twenty-nine are newly named by these authors from the New Albany. Nine more had previously been named from it or from its counterparts in other states, and three are left without specific names. This leaves but two species—*Kalymma grandis* and *Periastron reticulatum*—that occur elsewhere, and neither of these is a recognized horizon marker. In other words, thirty-eight of the forty named species are endemic to the New Albany shale. One cannot deny that there may be Devonian elements in the flora, but it is extremely doubtful that they provide the necessary data to prove that the Upper New Albany shale is of Devonian age. To try to use them as proof of Devonian age is very likely to lead to erroneous conclusions. If these Devonian elements existed by themselves, one would be little inclined to deny their apparent value, but there are other factors involved upon which I shall comment briefly.

On the Mississippian side of the ledger, we find among the plants listed by Read and Campbell the following typical Lower Carboniferous genera: *Lepidodendron*, *Lepidostrobus*, *Cladoxylon*, *Stenomyelon*, *Clepsydropsis*, *Calamopitys*, *Kalymma*, *Steloxylon*, *Asterocalamites*, and *Lyginorachis*. Some of these genera do extend downward into the Upper Devonian, but they are essentially Lower Carboniferous; and on the basis of any of them, or any combination of them, one would assume Carboniferous age. No flora of proved Devonian age anywhere contains all these. Therefore, it seems that, when all the evidence is brought together and the correlative value of all the plant types is appraised, there is at least as much in favor of Mississippian as of Devonian age, and the balance is, the author believes, decidedly on the side of the Mississippian. In fact, when we look at a list of petrified plants from the Cementstone group of the Calciferous sandstone of Great Britain we find four genera (*Lyginorachis*, *Calamopitys*, *Stenomyelon*, and *Cladoxylon*) in common with the New Albany shale flora. When these particular genera are found together, one can come to almost no other conclusion than that they are of Lower

Carboniferous age. Then there are still other genera—*Clepsydropsis* and *Protocalamites*, for example—that are decidedly more at home in the Lower Carboniferous than in the Devonian. I therefore concur with Campbell, who, in his latest paper, draws the Devonian-Mississippian boundary between the Blackiston and the Sanderson formations. In his analysis of the New Albany shale series, Campbell comments upon the uniformity of conditions of deposition which extended from Tully to Kinderhook time; and it is therefore not surprising to find a goodly number of ancient plant types held over from the preceding Devonian flora and to find them intermingled with later ones that come in toward the close of the long interval of black-shale deposition.

## REFERENCES CITED


BELL, W. A. (1929) Horton-Windsor district, Nova Scotia: Canada Geol. Survey Mem. 155.

CAMPBELL, G. (1946) New Albany shale: Geol. Soc. America Bull. 57, pp. 829–908.

CRIBBS, J. E. (1933) *Cordaites missouriense*, from the Lower Carboniferous of Missouri: Am. Jour. Botany, vol. 22, pp. 427–438.

——— (1938) A new fossil plant from the Reed Spring formation of southwestern Missouri: *ibid.*, vol. 25, pp. 311–321.

——— (1939) *Cauloxylon ambiguum*, gen. et sp. nov., a new fossil plant from the Reed Spring formation of southwestern Missouri: *ibid.*, vol. 26, pp. 440–449.

——— (1940) Structure of fossil stem of pityean affinity, from the Reed Spring formation of Missouri: Bot. Gazette, vol. 101, pp. 582–597.

JONGMANS, W. J., and GOTHAN, W. (1933) Florenfolge und vergleichende Stratigraphie des Carbons der östlichen Staaten Nord-Amerikas. Vergleich mit West-Europa: Geol. Bur. Heerlen, pp. 17–44.

———; ———; and DARRAH, W. C. (1937) Beiträge zur Kenntnis der Flora der Poconoschichten aus Pennsylvanien und Virginia: Cong. pour l'avancement études de stratigraphie carbonifère, Heerlen, 1935, Compte rendu, pp. 423–444.

READ, C. B. (1936) A Devonian flora from Kentucky: Jour. Paleontology, vol. 10, pp. 215–227.

———, and CAMPBELL, G. (1939) Preliminary account of the New Albany shale flora: Am. Midland Naturalist, vol. 21, pp. 435–453.

WHITE, D. (1913) The Horton flora, Guidebook I: Excursion in eastern Quebec and the Maritime Provinces, pt. 1, pp. 144–146.

——— (1926) General features of the Mississippian floras of the Appalachian trough: West Virginia Geol. Survey, Mercer, Monroe and Summers Counties, pp. 837–843.

——— (1937*a*) Fossil flora of the Wedington sandstone member of the Fayetteville shale: U.S. Geol. Survey Prof. Paper 186-B.

——— (1937*b*) Fossil plants from the Stanley and Jackfork sandstone in southeastern Oklahoma and western Arkansas: U.S. Geol. Survey Prof. Paper 186-C.

# PALEONTOLOGICAL FEATURES OF MISSISSIPPIAN ROCKS IN NORTH AMERICA AND EUROPE[1]

RAYMOND C. MOORE
University of Kansas

ABSTRACT

Comparison of marine fossil assemblages known from deposits of Mississippian age in North America and Europe shows many striking similarities in the nature and distribution of faunas but also brings out some strongly divergent characters. The occurrence of identical or nearly identical types of highly organized invertebrates, especially among the echinoderms and cephalopods, in correspondingly arranged successions of rather narrowly restricted zones affords a good basis for recognizing approximate stratigraphic equivalents in the two continents. Zonation of the European deposits is more detailed and better defined, however, than in most Mississippian sections of North America.

A conclusion of chief importance is that Mississippian deposits on both sides of the Atlantic are divisible in a similar manner into two main parts, classifiable as series. The Lower Mississippian rocks of North America, comprising Kinderhookian and Osagian beds, which have been called the "Waverlyan Series," correspond to the Tournaisian strata of Europe. Upper Mississippian formations, which are classed as belonging to the Meramecian and Chesterian portions of the American succession, collectively named the "Tennesseean Series," are judged on paleontological grounds to represent Viséan and lower Namurian deposits of the European continent. Distinctive upper Namurian marine fossils are unknown from North America, either in youngest Mississippian or in oldest Pennsylvanian beds; inasmuch as the Namurian seems everywhere to be an essentially conformable succession, no strongly marked break in sedimentation, such as marks the Mississippian-Pennsylvanian boundary in most parts of North America, is found at the summit of deposits of Mississippian age in Europe.

## INTRODUCTION

The purpose of this paper is to offer in compact form a comparison of paleontological data pertaining to Mississippian deposits of North America and Europe. As part of the discussion directed to general appraisal of our progress in working out problems of Mississippian stratigraphy in North America, the focus of my contribution to this symposium is intended to be on paleontological zonation of the American Mississippian formations rather than primarily on intercontinental correlation. From this viewpoint, interest in European fossils belonging to the Mississippian part of the column hinges mainly on the manner and extent of their helping us to proper evaluation of our paleontological record. We are not concerned especially about identifying the European equivalents of the Keokuk limestone or other parts of our succession, and we are not interested primarily in fixing the stratigraphic position of various British crinoid-bearing beds in terms of our richly fossiliferous crinoidal strata. The objectives in the present comparative study are to learn how closely parallel the Mississippian faunas on opposite sides of the Atlantic run in their evolution and to judge how much the zonation of European strata, which is much better defined than that in North America, may throw light on work yet to be done by us.

## ZONATION OF EUROPEAN LOWER CARBONIFEROUS FORMATIONS

European deposits, which on the basis of contained fossils are recognized as equivalent to Mississippian formations of North America, are somewhat narrowly classified according to paleontological zones, and these are identified both in the continental portion of western Europe and in the British Isles. Based on sections in Belgium, the strata classed as belong-

[1] Manuscript received January 31, 1948.

ing to the Lower Carboniferous in Continental Europe are commonly called "Dinantian," and those defined as Upper Carboniferous (or by some geologists as Middle Carboniferous) begin with formations referred to the Namurian division, which also is named from outcrops in Belgium. Dinantian deposits are divided into Tournaisian beds below and Viséan strata above, which are widely recognized through Eurasia and northern Africa. Below the base of the Tournaisian, and included by some geologists in it, is a relatively thin stratigraphic division variously called "Etroeungtian" or "Strunian." The age of beds assigned to this division has been debated, and there is a tendency to regard it as a division distinct from Tournaisian. The presence of Upper Devonian elements, especially clymenid ammonoids, in Etroeungtian deposits supports their assignment to uppermost Devonian (Famennian), but there are also Carboniferous types of invertebrates. Consensus assigns the Etroeungtian to the base of the Carboniferous. It is noteworthy that the stratigraphic succession in Belgium and western Germany from Devonian into Carboniferous beds is conformable. Likewise, there is absence of a pronounced stratigraphic break at the top of the Dinantian, where a boundary is drawn between Viséan and Namurian beds.

In southwestern England, beds classed as Lower Carboniferous have been termed "Avonian" because of presumed slight difference in stratigraphic span of the section there, as compared with the type Dinantian; elsewhere in the British Isles they are commonly known simply as "Carboniferous limestones." The type Avonian section is along the Avon River near Bristol, and in this area a series of zones, delimited by means of corals and brachiopods, was defined some forty years ago. These zones, designated by index letters of the guide genera (K, Z, C, S, D) have come to be standard for reference of sections in other parts of the British Isles (fig. 1). Equivalents of Etroeungtian strata of Continental Europe are recognized as belonging to the K zone and are included in the lowermost Avonian. Near Bristol, the K beds rest conformably on upper Old Red sandstone of Devonian age. It is now recognized that the line of division corresponding to the boundary between Tournaisian and Viséan strata falls in the middle part of the C zone of the British succession, that is, between subzones called $C_1$ and $C_2$ (fig. 1). Overlying the Avonian or Carboniferous limestones comes the Millstone grit, which is classed as the lower main division of the Upper Carboniferous sequence. The dominantly clastic deposits of the Millstone grit are recognized as equivalent to the Namurian of Continental Europe.

The upper part of Dinantian beds in Germany, Belgium, and northern England is composed mainly of dark shale and siliceous limestone, constituting the so-called "Culm facies" (Kohlenkalk). The strata of this facies lack the corals and most of the brachiopod species that are used to zone the Avonian deposits, but they contain pelecypods, particularly *Posidonia*, and several sorts of ammonoid cephalopods. In northern England, Upper Dinantian zones, defined by means of mollusks, are overlain conformably by ammonoid zones of the Namurian sequence, and these are designated by index letters (B, P, E, H, R, G) of selected guide genera (figs. 1, 15, 16). The problem of correlating the coral-brachiopod zones and the ammonoid zones belonging to different facies was not easily or immediately solved; but the finding of interfingering relationships be-

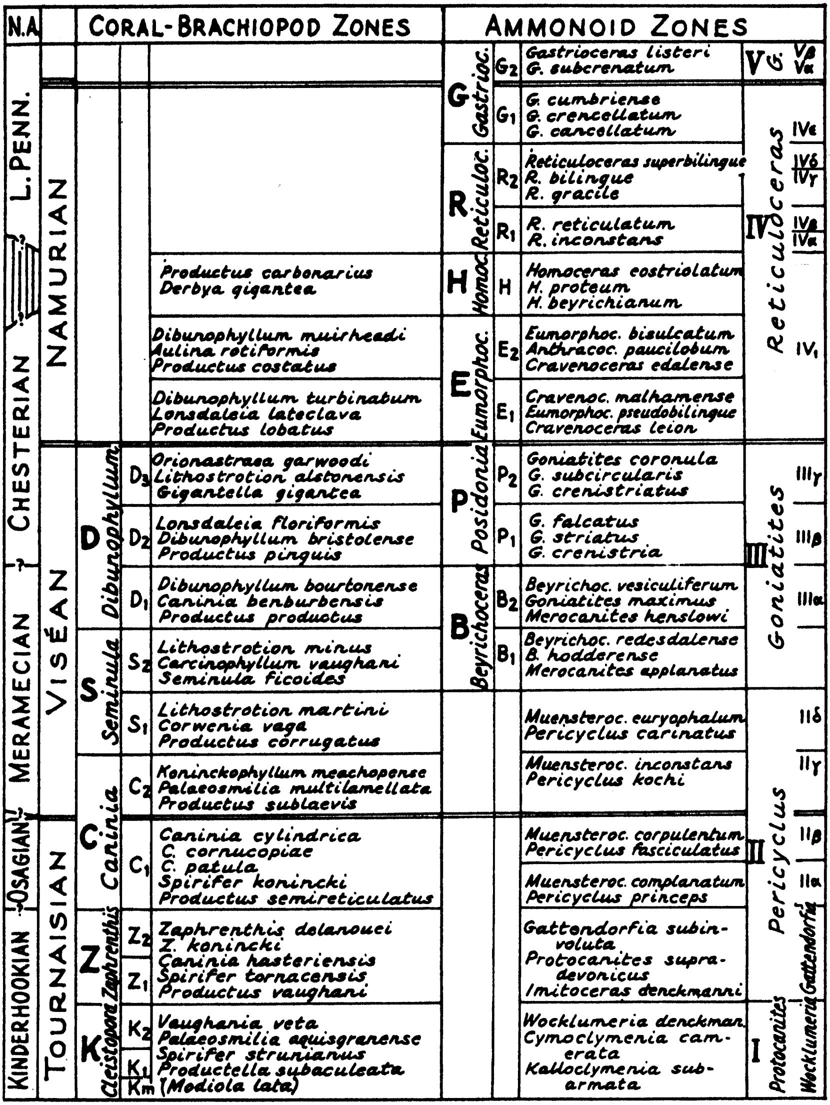



FIG. 1.—Zonation of Lower Carboniferous rocks in western Europe, chiefly the British Isles, showing (*at left*) inferred correlation with American Mississippian divisions. The column at right indicates ammonoid zones, defined in Germany by H. Schmidt.

tween the two sets of deposits has now established correlation of the B zone with the $S_2$ and $D_1$ coral-brachiopod subzones, and the P zone with the $D_2$ and $D_3$ coral-brachiopod subzones. Corals and brachiopods corresponding to the E and H zones of the Millstone grit have been recognized, but these have not been defined as separately lettered zones (fig. 1).

Ammonoid zones in western Germany have been defined in a slightly different manner by Schmidt (1928) and by Hudson and Turner (1933*b*). In upward order, zones of *Protocanites*, *Pericyclus*, *Glyphioceras* [*Goniatites*], and *Reticuloceras* were designated by the Roman numerals I–IV, and these have been subdivided by Greek letters into subzones (fig. 1).

## ZONATION OF AMERICAN MISSISSIPPIAN ROCKS

Mississippian deposits of North America have not been zoned comprehensively on the basis of invertebrate or other fossils. Also, there is far from complete agreement as to stratigraphic correlation of sections having different sedimentary facies occurring in various parts of the continent. Some broadly significant paleontological characteristics have long been recognized, however. Chief among these are the recognized dominance of a host of camerate crinoids in Lower Mississippian strata, in contrast to the virtual absence of these fossils in later Mississippian beds, and the abundance of pentremitid blastoids and the common occurrence of compound corals belonging to *Lithostrotion* in the upper part of the Mississippian succession. In the central Mississippi Valley, remains of the distinctive crinoid genera *Talarocrinus* and *Pterotocrinus* characterize higher Mississippian (Chesterian) strata, whereas these genera are entirely unknown in older rocks.

An effort to define faunal zones in the standard Mississippian section was published by Stuart Weller (1926) on the basis of extensive field studies made by him in Iowa, Illinois, Missouri, and Kentucky (fig. 2). He recognized fourteen zones, among which five were named in terms of brachiopods, one on a coral, and others by combinations of species representing bryozoans, crinoids, brachiopods, blastoids, gastropods, and pelecypods. The indicated stratigraphic distribution of the zones, according to Weller (1926, p. 324), was Kinderhookian, one zone; Osagian, two zones; Meramecian, four zones; and Chesterian, seven zones. Present opinion is that a majority of the zones defined by Weller are not widely applicable to classification and correlation of Mississippian deposits in North America.

Work on Kinderhookian rocks of Iowa has resulted in differentiation of eighteen paleontological zones (Laudon, 1931) (fig. 2). Some of these are known to be applicable in neighboring states, but the geographical distribution and usability of most of these zones is undetermined. Deposits of Kinderhookian age are seemingly many times thicker in the Rocky Mountain region of western United States and Canada than they are in the Mississippi Valley area. Some of the zone fossils discriminated in Iowa are known to occur in the western Kinderhookian (most of the Madison and equivalent beds).

Studies by Laudon (1947, p. 1161) of lower Osagian rocks in Missouri and Iowa have led him to define seven zones in the Burlington limestone on the basis of camerate crinoids and of blastoids (fig. 2). These zones were found to be clearly differentiated and recognizable along most of the outcrops and had been found applicable in northeastern, central,

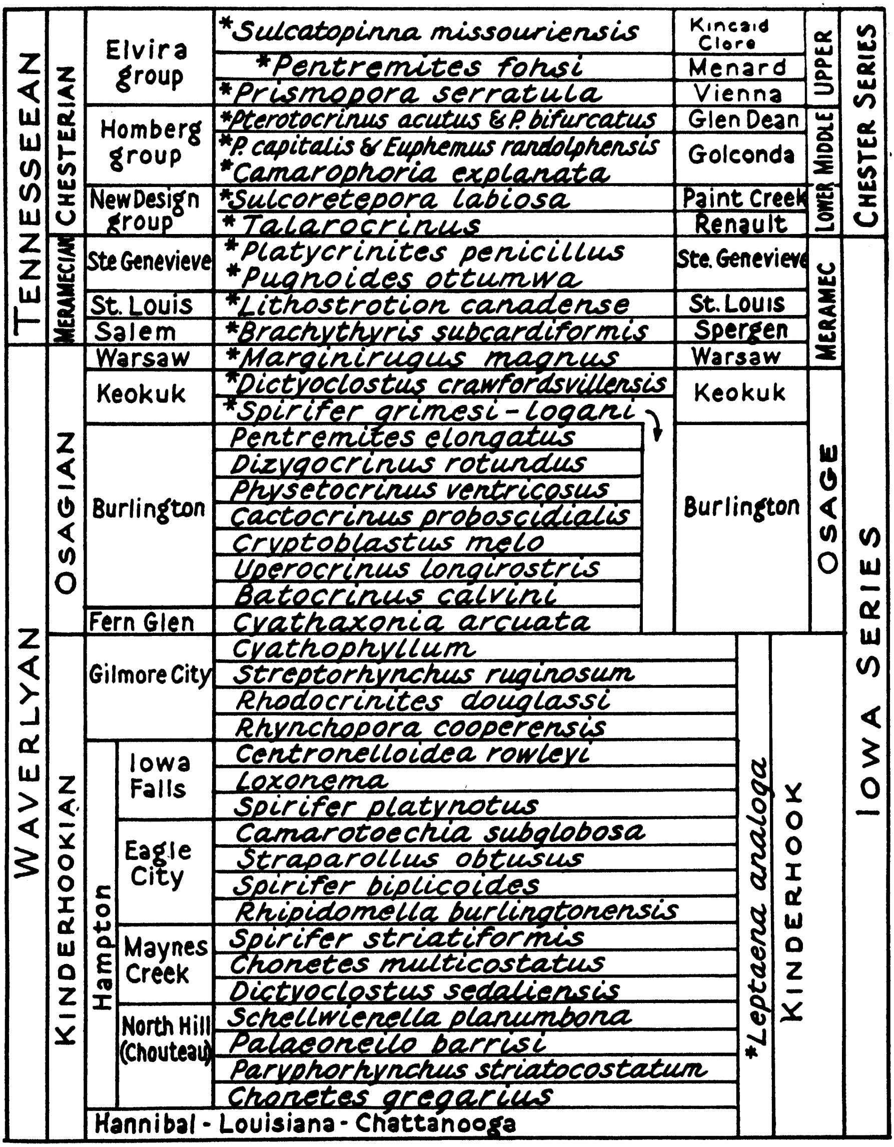



FIG. 2.—Paleontological zones in American Mississippian rocks. The stratigraphic classifications shown at right and the zones indicated by asterisks are according to Weller (1926). The stratigraphic classification at left is based on Van Tuyl, Laudon, and Moore. Paleontological zones below the Keokuk are after Laudon (1931, 1933) (except *Cyathaxonia arcuata*).

and southwestern Missouri. It is believed that other parts of the Mississippian section can be very satisfactorily zoned in terms of echinoderms; but published work along this line has not appeared.

Miller and Furnish (1940, p. 357) have suggested the placement of American Mississippian ammonoid-bearing beds in terms of European zones, but they have not defined or named such zones for the classification of the American section. They suggest that the middle Viséan *Beyrichoceras* zone, and perhaps most of the *Posidonomya* [*Posidonia*] zone of the upper Viséan of Europe, are equivalent to Meramecian deposits in North America, whereas the lower Namurian *Eumorphoceras* zone is recognized as Chesterian.

## COMPARISON OF BIOLOGICALLY DEFINED GROUPS OF MISSISSIPPIAN INVERTEBRATE FOSSILS FROM NORTH AMERICA AND EUROPE

### GENERAL STATEMENT

The following portion of this paper is devoted to the comparison of several biologically defined groups of Mississippian fossils, but it is recognized as decidedly incomplete, even though some of the tabulations presented are reasonably comprehensive. Available time has permitted organization of data sufficient only to support somewhat qualified conclusions. In the groups mainly studied, the nature of evidence now available is believed to be appraised with reasonable adequacy. A major difficulty has been the satisfactory determination of the stratigraphic placement of described fossils. Mostly, these data can be determined with assurance, but, because of inadequacy of records or lack of information as to stratigraphic correlations, especially among European records, some stratigraphic assignments are insecure. A factor that also affects part of the comparisons offered is correctness of generic identification. This means that among somewhat loosely defined genera, such as *Poteriocrinites*, *Pachylocrinus*, and *Actinocrinites*, for example, records of occurrence on opposite sides of the Atlantic probably fail to take account of actual distinctions of generic importance. The recorded distribution of many narrowly defined genera, like *Hypselocrinus* and *Ainacrinus*, which have been differentiated, respectively, in North America and Europe and are not now known to be represented by species on the other continent, is likely to be modified by future studies. In spite of these difficulties, many features of the comparisons are judged to be worthy of special notice.

### FORAMINIFERS

Foraminiferal remains are fairly common in some Mississippian deposits, but they have not been studied extensively, and there is no present indication that they may be very useful for purposes of paleontological zonation. The genus *Endothyra*, which is common, especially in oölitic limestones such as the Salem and Ste. Genevieve, was originally described from Lower Carboniferous rocks of England, probably from beds of Viséan age. Studies by Edward Zeller, of the University of Kansas, have shown that shells referable to *Endothyra* are distributed from low in the Kinderhookian part of the American Mississippian section to Chesterian and that species having value for stratigraphic differentiation may be recognized. Liebus (1931) has described 58 species of Lower Carboniferous foraminifers from Germany; these are assigned to 26 genera. Included are species of *Endothyra* and forms classed as *Fusulinella*, the latter being almost cer-

tainly not *Fusulinella* but referable to *Millerella*. Examples of *Millerella* from Upper Mississippian rocks of North America have been distinguished by Zeller.

CORALS

Corals play a prominent part in the paleontological zonation of the Lower Carboniferous rocks in northwestern Europe, especially the coral-brachiopod facies of the British Isles. Hill (1938) has recognized three distinct facies faunas among the corals, which recur wherever a proper environment was established. These faunas are (1) a cyathaxonid fauna, composed largely of small solitary corals lacking dissepiments, occurring most commonly in shaly deposits; (2) caninid-clisiophyllid faunas, characterized by large solitary forms having dissepiments, occuring generally in light-colored well-bedded limestones; and (3) reef coral faunas, characterized by compound corals having dissepiments. The cyathaxonid faunas especially distinguish the rocks of Tournaisian age but extend into higher strata, whereas the other two assemblages are almost entirely developed in strata of Viséan and Namurian age.

Distribution of the most important coral species recognized in the British area is plotted in accompanying charts, in which forms reported by Hill are rearranged according to stratigraphic distribution, short-ranging types in one group (fig. 3) and long-ranging types in the other (fig. 4). Characteristic corals are illustrated in figure 5. The seeming distinctness of stratigraphic differentiation in terms of corals is well indicated. It is noteworthy that Hill has partly modified the zonal arrangement indicated by the generally recognized lettered subzones of the upper Viséan and has extended the zones to include lower Namurian deposits. Her Zone 1 is the same as $D_1$; it is the first reef-coral fauna, characterized by *Lithostrotion martini*, *L. minus*, *L. junceum*, *L. irregulare*, and *Lonsdaleia duplicata*. Coral Zone 2 comprises the $D_2$ subzone and about half the $D_3$ subzone (upward to the horizon of the Singlepost limestone). It is characterized by the presence of *Corwenia*, *Aulina furcata*, and the appearance of *Orionastraea* and *Palaeosmilia regia*. Coral Zone 3 comprises upper beds of $D_3$ and possibly the entire $E_1$ subzone; it is especially marked by the occurrence of *Lonsdaleia alstonensis*, *L. laticlavia*, and several other forms. The fourth coral zone belongs entirely in the lower Namurian, being made up mainly of the $E_2$ subzone; it is marked by *Aulina senex* and *A. rotiformis*, associated with forms of *Lithostrotion*.

Mississippian coral faunas from North America are very inadequately known, chief work during recent years having been done by Easton (1943*a*, *b;* 1944; 1945*a*, *b*, *c*); Kelly (1942), Merriam (1942), Jeffords (1943), and Sloss (1945). A compilation of the reported stratigraphic occurrence of some 70 coral genera reported from Mississippian rocks of North America and Europe is given in figure 6. Among these genera, more than 25 per cent occur on both sides of the Atlantic, and, in general, they have corresponding stratigraphic placement. Some forms, such as *Amplexus*, have relatively little meaning in the light of the present understanding of the polyphyletic nature of the types included in the group and the lack of basis for recognizing significant short-ranged forms. Other genera, particularly *Lithostrotion*, *Lonsdaleia*, *Diphyphyllum*, and *Dibunophyllum*, seemingly have much stratigraphic importance. These are almost exclusively forms characteristic of the

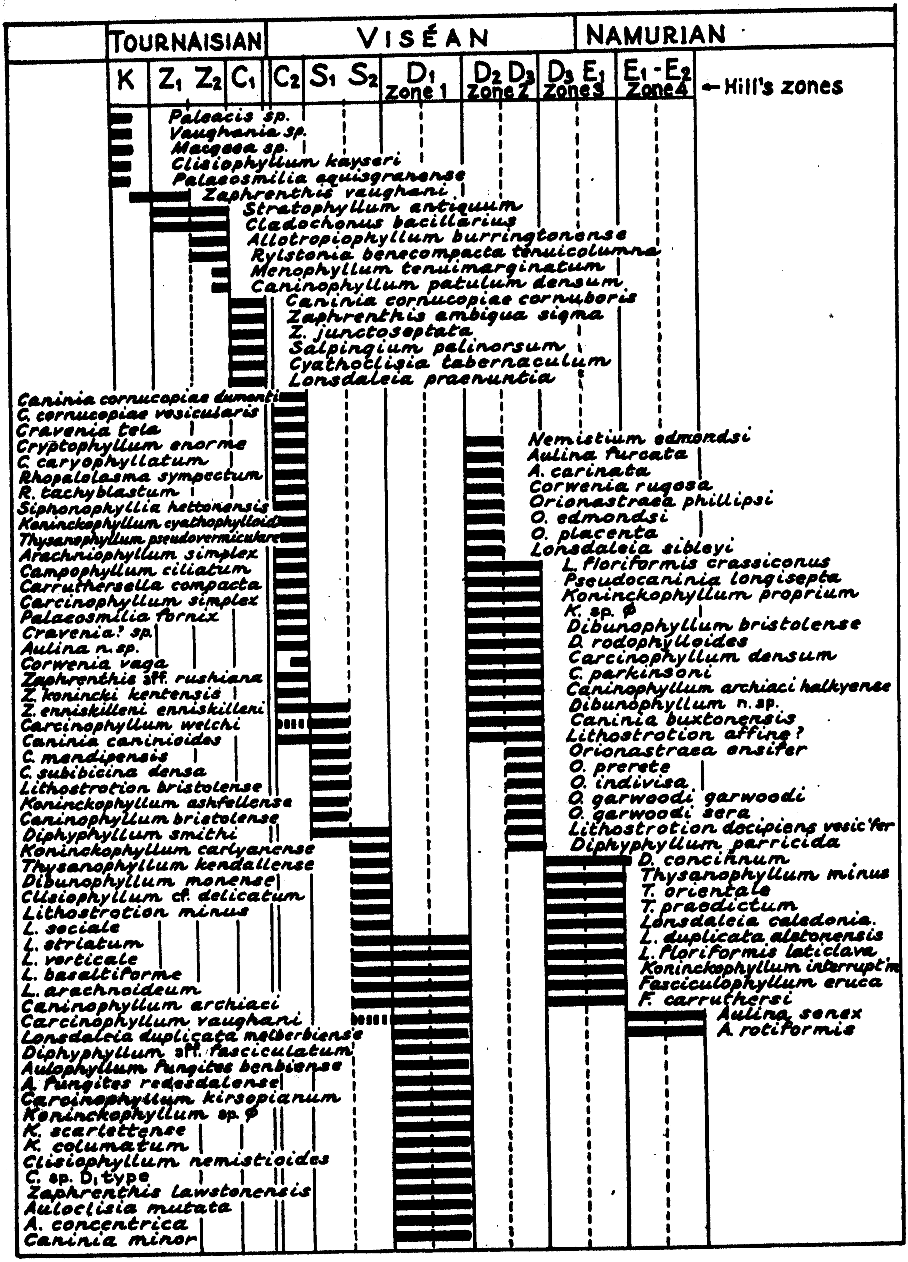



FIG. 3.—Short-range coral species of British Isles, arranged according to paleontological zones (data from Dorothy Hill, 1938).

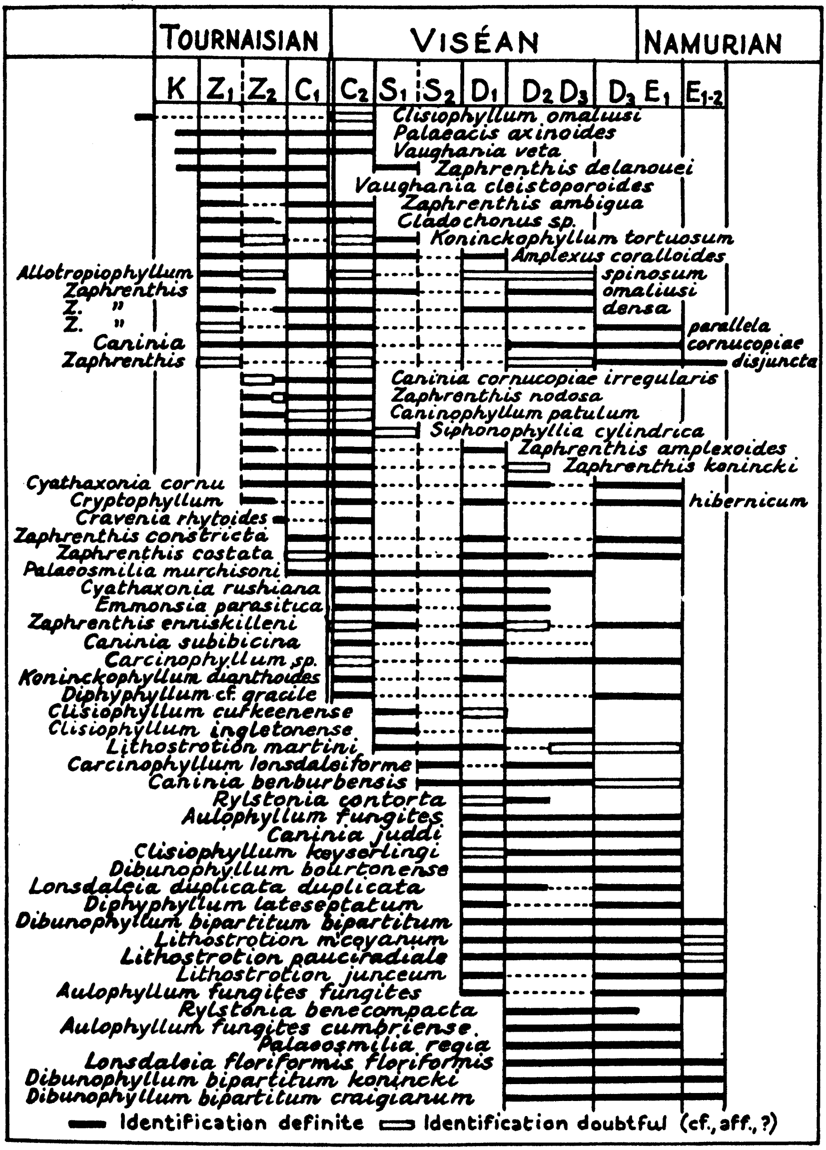



FIG. 4.—Long-ranging coral species of the British Isles, arranged by paleontological zones (data from Dorothy Hill, 1938.


FIG. 5—Representative Lower Carboniferous coral species, chiefly from the British Isles and Belgium (somewhat reduced).

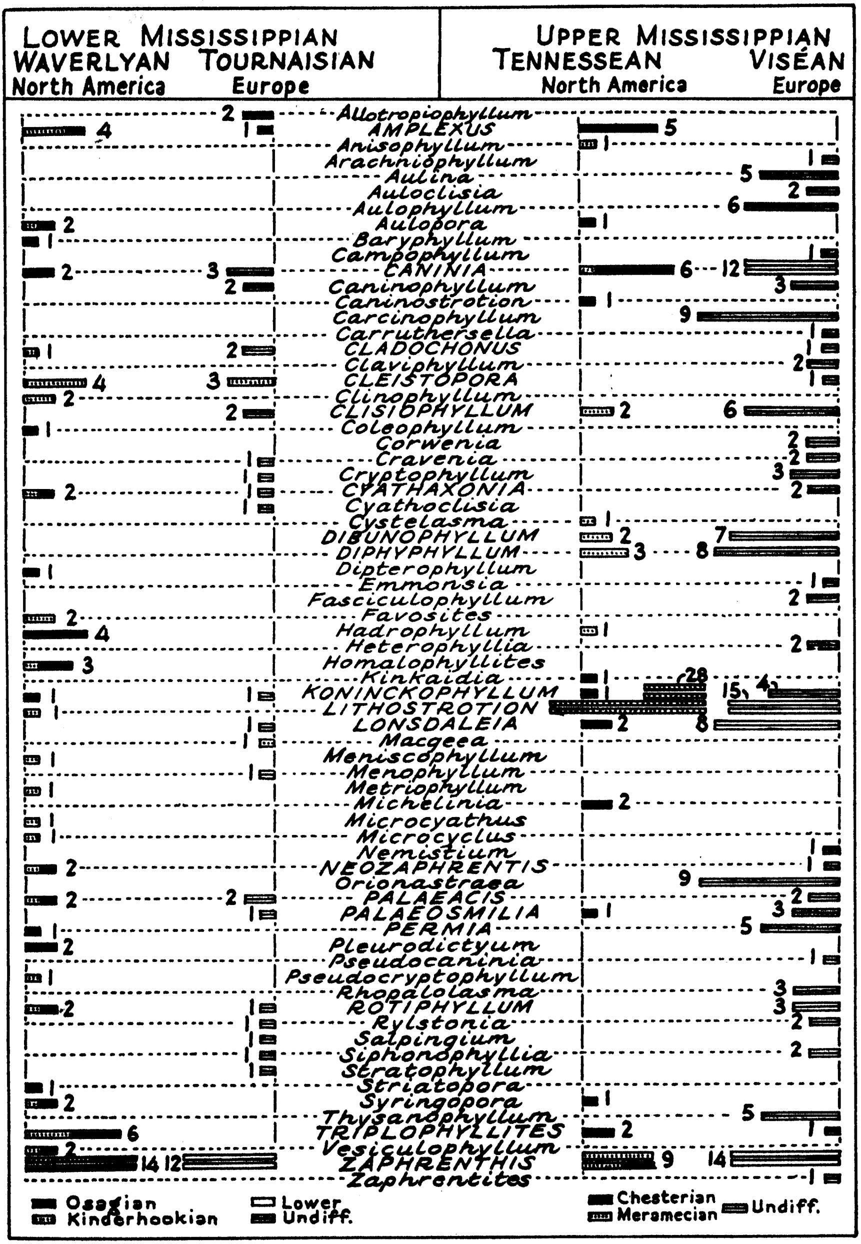



FIG. 6.—Comparison of the stratigraphic distribution of coral species recognized in rocks of Mississippian age in North America and Europe. Genera represented in both continents are indicated by capital letters.

Tennessean or Upper Mississippian Series in North America and found in Viséan and lower Namurian rocks of northwestern Europe. *Cleistopora*, on the other hand, is represented both in North America and in Europe in Lower Mississippian strata.

In the Canadian Rockies north of Jasper Park, thick Kinderhookian limestone containing *Leptaena analoga* and other characteristic brachiopods, associated with crinoids and blastoids, is characterized by extraordinary abundance of zaphrenthid corals, suggestive of the Z zone of the British Isles (L. R. Laudon, personal communication). The genus *Caninia*, for which the C zone, straddling the Tournaisian-Viséan boundary in the British area, is named, occurs both in Waverlyan and Tennesseean rocks of North America. A species of *Caninia* (*C. arcuata*) described by Jeffords (1943, p. 548) from lower Osagian rocks of New Mexico is reported to resemble very closely *C. cornucopiae* from upper Tournaisian beds of Belgium and England. Sloss (1945, p. 309) reports *Caninia cornucopiae* and *Lithostrotion* cf. *L. irregulare* from the Yakinikak limestone of northwestern Montana. *Caninia cornucopiae* ranges from $Z_1$ through $C_2$ of the British Mississippian zones, being most abundant in $C_1$; but closely similar corals are recorded from upper Viséan and lower Namurian strata of England (Hill, 1938, p. 6). *Lithostrotion* is almost wholly confined to Viséan rocks of Europe; *L. irregulare* is common in the upper Viséan. In North America this genus, including *Lithostrotionella* and *Siphonodendron*, is now represented by some 29 species (fig. 6), all but one occurring in Meramecian or younger rocks. The occurrence of two species of *Lithostrotion*, accompanied by *Dibunophyllum*, *Gigantella*, and *Striatifera*, in Mississippian rocks of the Pacific border (Merriam, 1942), indicates close correspondence with Viséan rocks of the British Isles, inasmuch as all these genera are characteristic of the high Mississippian rocks in western Europe.

### BLASTOIDS

The echinoderms generally are judged to have much importance for faunal comparison of deposits of Mississippian age on opposite sides of the Atlantic. This is because of the relatively high structural organization and distinctive morphologic features among various groups and especially because of relative abundance of these organisms in many places. Therefore, a survey of the several kinds of echinoderms and their stratigraphic distribution merits special attention.

Among blastoids 19 genera, distributed among 8 families, are recognized in Mississippian rocks of North America and Europe. Eight genera are now known from America but have not been identified in western Europe; 5 are European genera, unrecognized on this side of the Atlantic; and there are 6 genera in common—*Codaster*, *Orophocrinus*, *Phaenoschisma*, *Mesoblastus*, *Orbitremites*, and *Pentremites* (fig. 7). Excepting *Orophocrinus* and *Pentremites*, species belonging to genera represented in both continents seem to be more common in Upper Mississippian deposits. In North America, *Codaster*, *Orophocrinus*, and *Orbitremites* are dominantly Lower Mississippian genera. *Pentremites* occurs in Lower Mississippian rocks on both sides of the Atlantic, but in North America it is mostly an Upper Mississippian form (67 species), whereas no representatives are known to occur in equivalent rocks of Europe. This is a striking contrast. *Schizoblastus*, *Carpenteroblastus*, and *Cryptoblastus*, which are important Lower Mississippian forms of North America, are unknown in Europe.

EDRIOASTEROIDS

Edrioasteroids are a relatively unimportant group of echinoderms, viewed from a stratigraphical standpoint, perhaps mainly because of their comparative rarity. One genus (*Lepidodiscus*) has been found in both North America and Europe, but in the former it occurs in Lower Mississippian rocks and in the latter in Viséan deposits (fig. 7).

CRINOIDS

The Mississippian Period was especially characterized by the abundance and extraordinary variety of crinoids, among which all three subclasses—Inadunata, Flexibilia, and Camerata—are well represented. A study of the distribution of genera and species of Mississippian crinoids is summarized in figures 8, 9, 10, and 12.

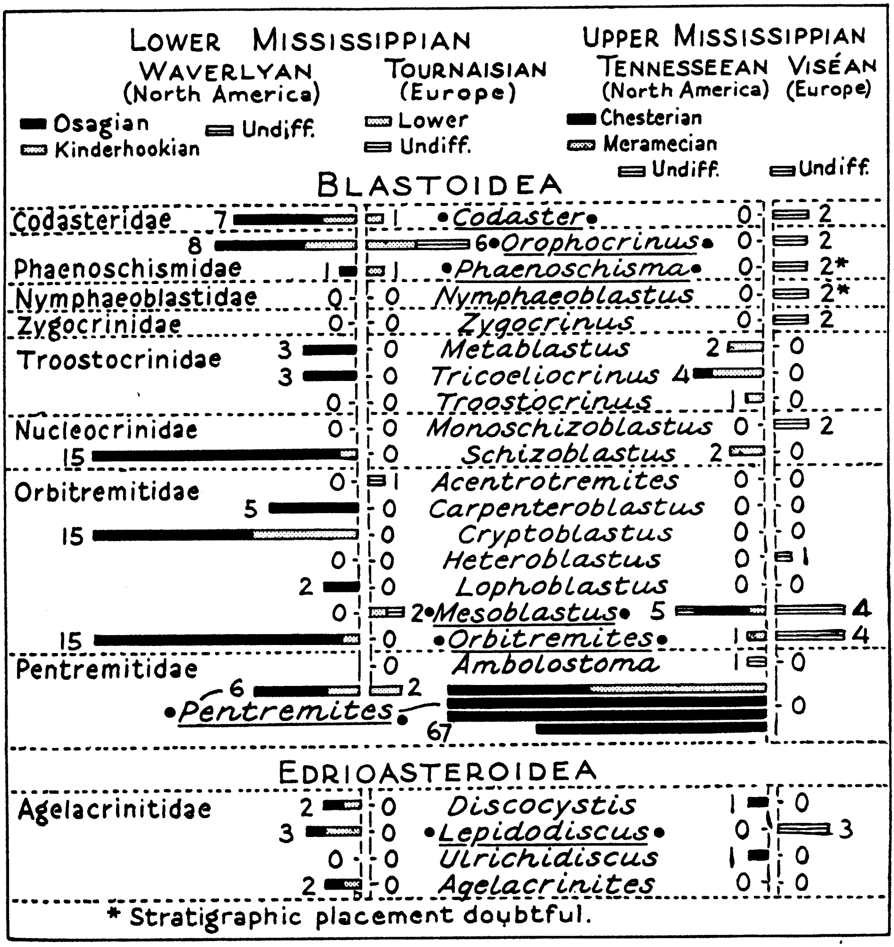



FIG. 7.—Comparison of Mississippian blastoids and edrioasteroids of North America and Europe. Genera common to both continents are underlined.

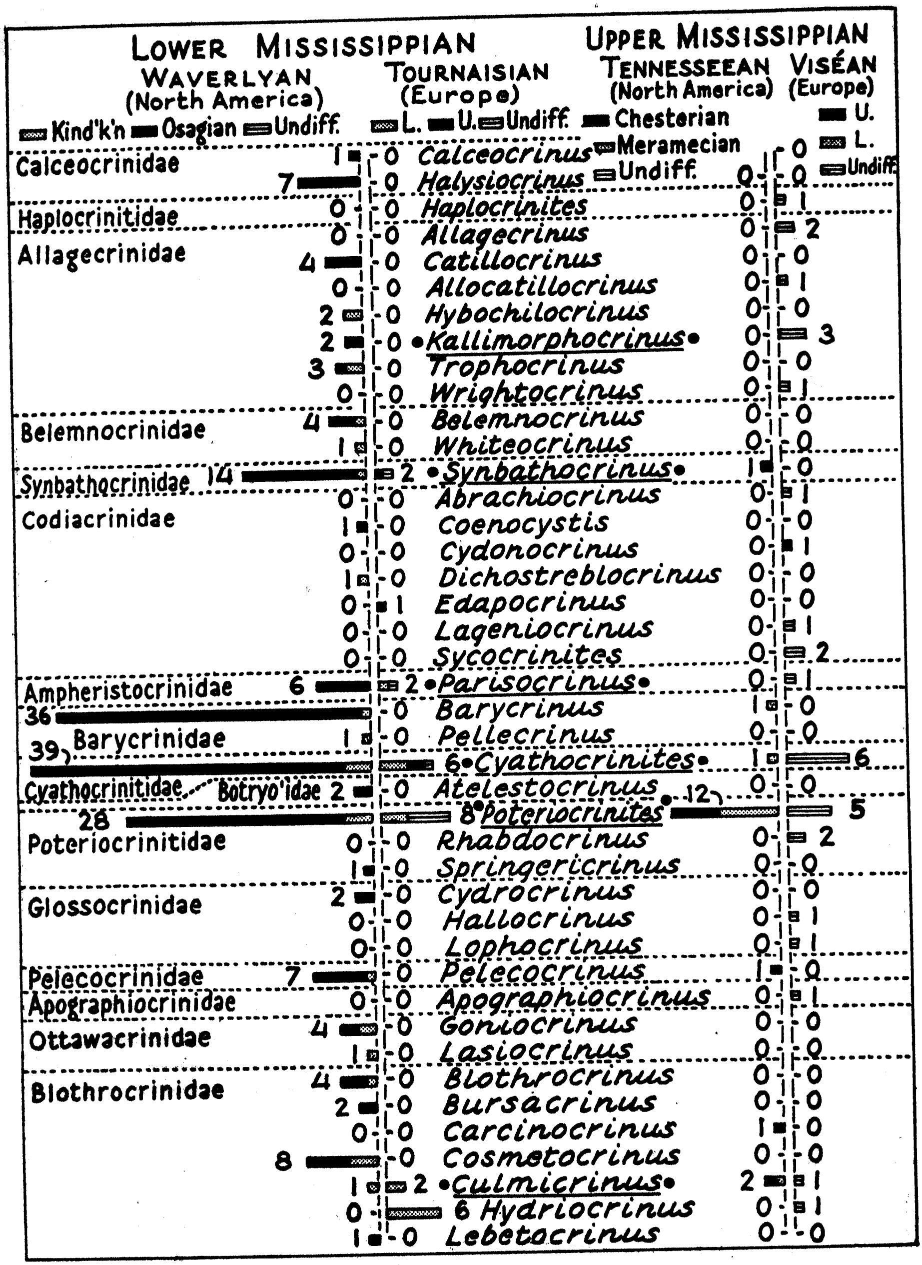



FIG. 8.—Comparison of Mississippian inadunate crinoids of North America and Europe. Genera common to both continents are underlined.

The very wide range of forms among inadunate crinoids is indicated by the numerous genera that have been recognized, and, as study continues, it is certain that many additional forms will be differentiated. The careful discrimination which is being employed in the definition of most genera signifies that the discovery of forms belonging to the same genus in North America and Europe furnishes reliable indication of a connection that may be interpreted to mean the existence of means for intermigration and probably close equivalence in geologic age. Conclusions as to age equivalence, however, are much less safely based on species belonging to genera of generalized form or simple structure, such as *Kallimorphocrinus* and *Synbathocrinus*, and probably such genera as *Cyathocrinites* and *Poteriocrinites*, as commonly identified, than in cases of many other inadunate genera having distinctive structural characters of cup and arms.

By and large, a survey of the inadunates does not indicate specially important points of similarity or discordance (figs. 8–10). It is of interest that the very distinctive, peculiarly specialized calceocrinids, especially *Halysiocrinus*, which are not uncommon in Lower Mississippian rocks of North America, have not been found in Europe. A distinctive but relatively unspecialized crinoid, *Barycrinus*, is found to be represented by 36 Waverlyan species in North America and in Europe. Distinctive inadunates that have importance in characterizing Lower Mississippian rocks of North America include *Catillocrinus*, *Belemnocrinus*, *Pelecocrinus*, *Blothrocrinus*, *Cosmetocrinus*, *Cercidocrinus*, *Coeliocrinus*, *Eratocrinus*, *Sarocrinus*, *Histocrinus*, and *Hypselocrinus*, none of which has yet been found in Europe. Similarly, diagnostic types of Upper Mississippian crinoids from North America that are unknown in Europe include *Dasciocrinus*, *Dinotocrinus*, *Rhopocrinus*, *Tholocrinus*, *Phacelocrinus*, *Eupachycrinus*, *Agassizocrinus*, and *Anartiocrinus*. Distinctive European inadunates occurring in Viséan strata, mainly in Scotland, include *Woodocrinus*, *Hydreionocrinus*, *Anemetocrinus*, *Aphelecrinus*, and *Scotiacrinus* (fig. 11); none of these genera has yet been identified west of the Atlantic.

Among common inadunate genera, probably the most significant are *Parisocrinus*, *Culmicrinus*, *Pachylocrinus*, *Scytalocrinus*, *Zeacrinites*, and *Phanocrinus*. The last-named two genera are especially characteristic of Upper Mississippian deposits. The genotype of *Graphiocrinus* is a Tournaisian species from Belgium and the British area; it is interesting to note that 14 species assigned to this genus occur in Kinderhookian and Osagian rocks of North America. No species of *Graphiocrinus* has yet been described from Upper Mississippian rocks, but the genus is represented by Pennsylvanian and possibly Permian forms from North America and Timor.

Among flexible crinoids, 17 genera of Mississippian forms have been described, and, of these, 6 are common to North America and Europe (fig. 10). These common genera include highly organized and distinctive forms, such as *Onychocrinus*, *Taxocrinus*, *Forbesiocrinus*, *Euryocrinus*, and *Wachsmuthicrinus*. Two genera of flexible crinoids—*Amphicrinus* and *Talanterocrinus*—described from western Europe but not known to occur in Mississippian rocks of North America, are definitely identified in Pennsylvanian rocks of this continent. The two lecanocrinid species referred to "*Stemmatocrinus*" (fig. 10) are from Osagian rocks of central Tennessee; they have been erroneously referred to an inadunate genus.

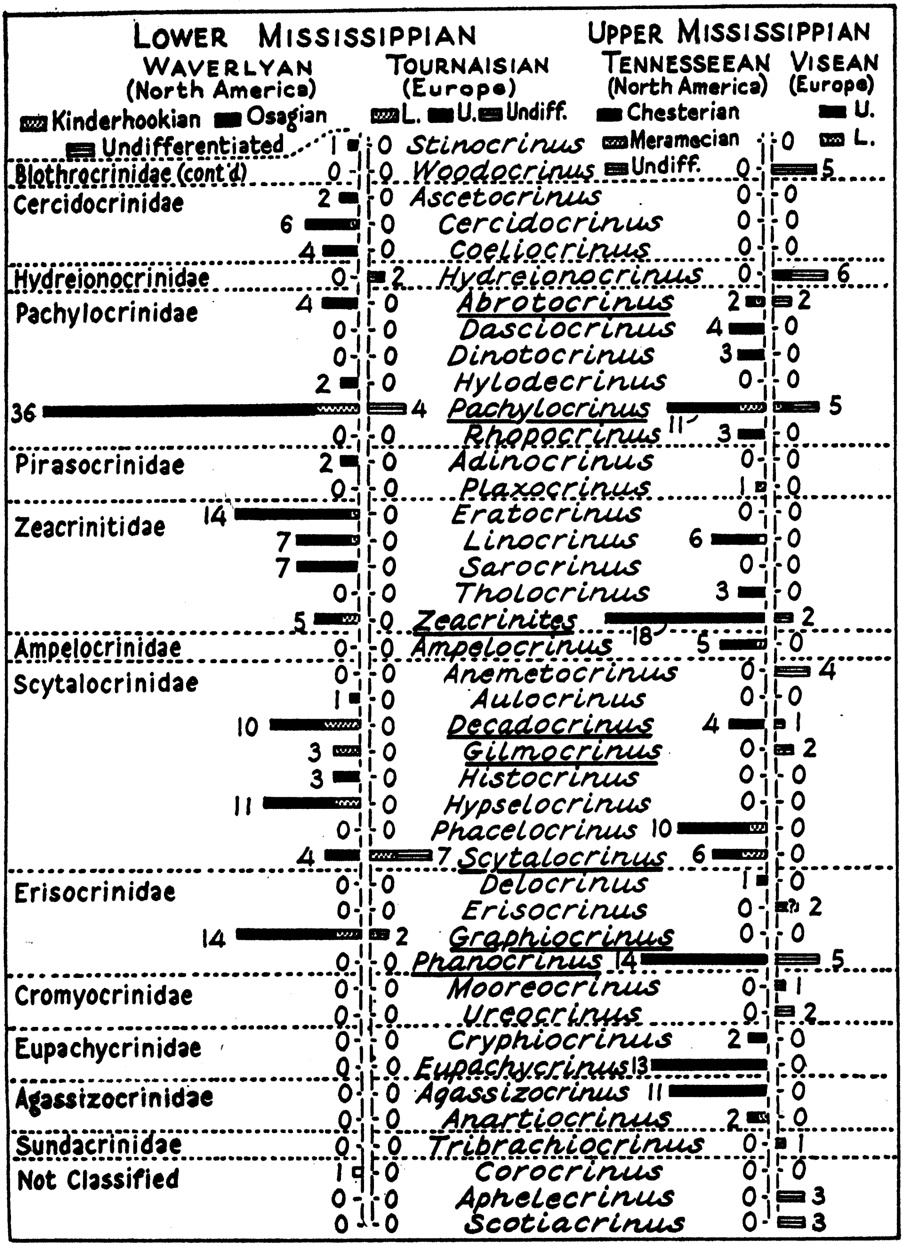



FIG. 9.—Comparison of Mississippian inadunate crinoids of North America and Europe (*continued*). Genera common to both continents are underlined.

In paleontological study of Mississippian rocks no group of fossils holds precedence over the camerate crinoids (figs. 11–12). This assertion does not rest on judgment that the camerates include the most highly organized crinoids, although forms are widely distributed geographically. Accordingly, the record of the camerates, as summarized in figure 12, is especially interesting in making a comparison of the Mississippian rocks of North America and Europe.

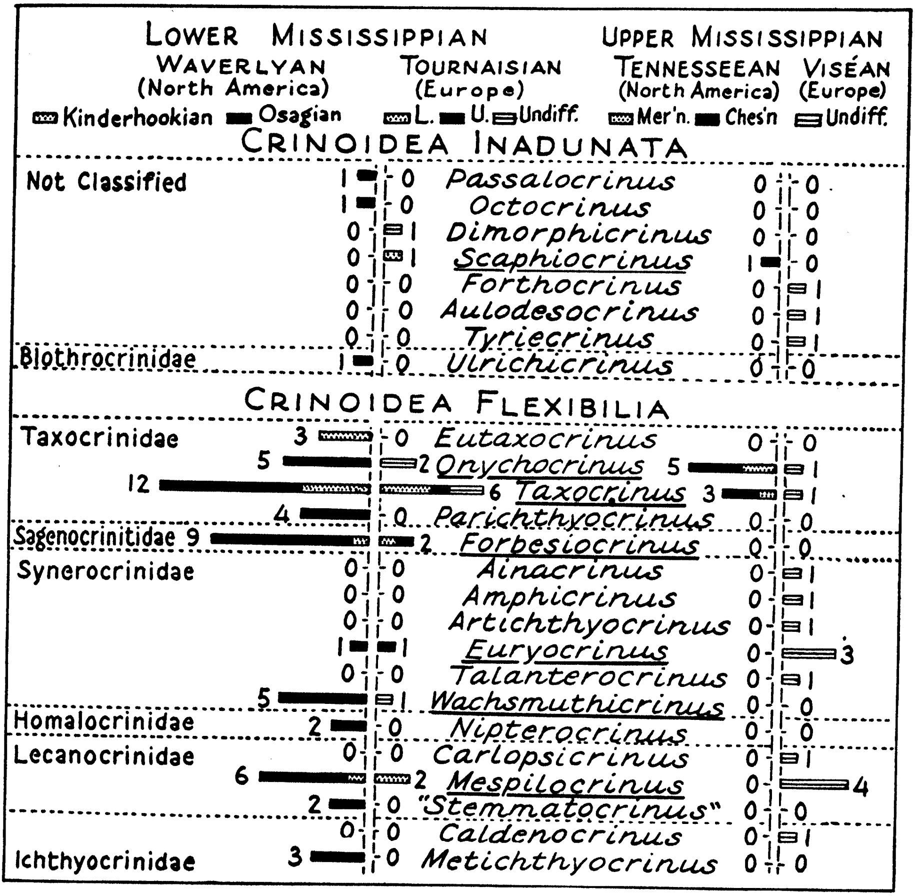



FIG. 10.—Comparison of Mississippian inadunate (*continued*) and flexible crinoids from North America and Europe. Genera common to both continents are underlined.

none have more complex structural elements; it contemplates mainly the amazing profusion and variety of these fossils, combined with observation of an extremely short stratigraphic range of almost all known species. Also, many

Among 35 now recorded Mississippian camerate genera, one-third are common to North America and Europe. The phylogeny of this division of the Crinoidea is not well understood, and it is almost certain that future research will

multiply the number of genera that are properly definable from these rocks which were formed at the time of the culmination of the camerate crinoid stock. Such new genera will be defined mainly by revisions of present known crinoids rather than by discovery of new forms.

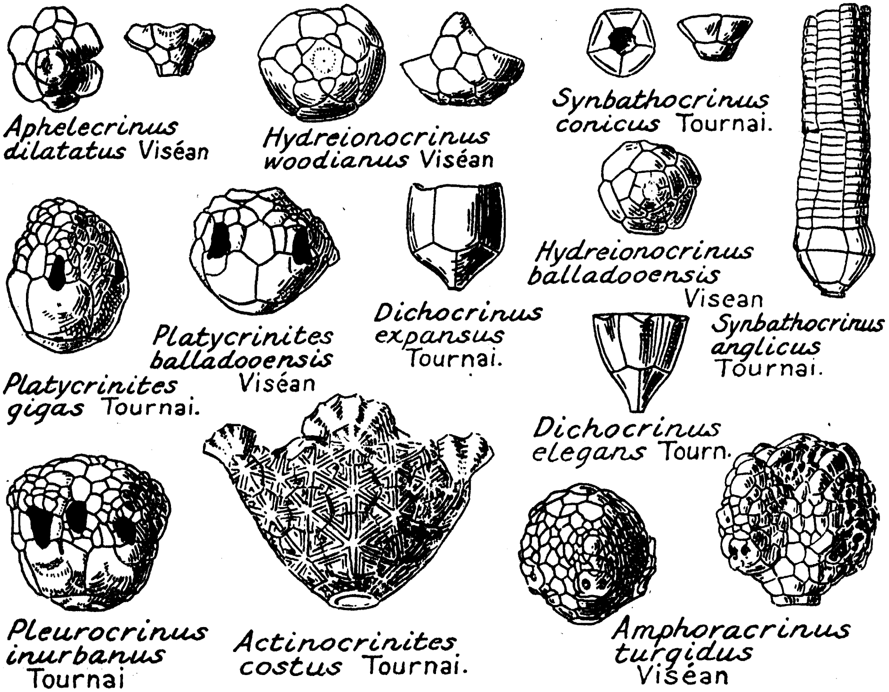



FIG. 11.—Representative inadunate and camerate crinoids from Lower Carboniferous rocks of the British Isles and Belgium (somewhat reduced).

The camerate crinoids now known from both sides of the Atlantic may be divided into four groups. The first comprises the rhodocrinitids, which began in the Ordovician and died out in Mississippian time; here belong *Rhodocrinites* and *Gilbertsocrinus*, which are wholly Lower Mississippian genera as known in North America but are present in both Tournaisian and Viséan strata of western Europe. The second group comprises the periechocrinitids and their derivatives classed in the families Periechocrinitidae, Actinocrinitidae, and Amphoracrinidae; this stock began in the Silurian Period, and some representatives of it persisted into the Permian. American and European genera in this group include *Megistocrinus*, *Amphoracrinus*, *Actinocrinites*, *Cactocrinus*, and *Physetocrinus*, all of which are exclusively Lower Missisippian forms on this continent but are represented in part by Viséan species in Europe. A third group that includes *Dichocrinus* and *Camptocrinus* belongs to the family Dichocrinidae. This is a relatively long-ranging stock, which has its largest differentiation in Lower Mississippian deposits but occurs in both continents in

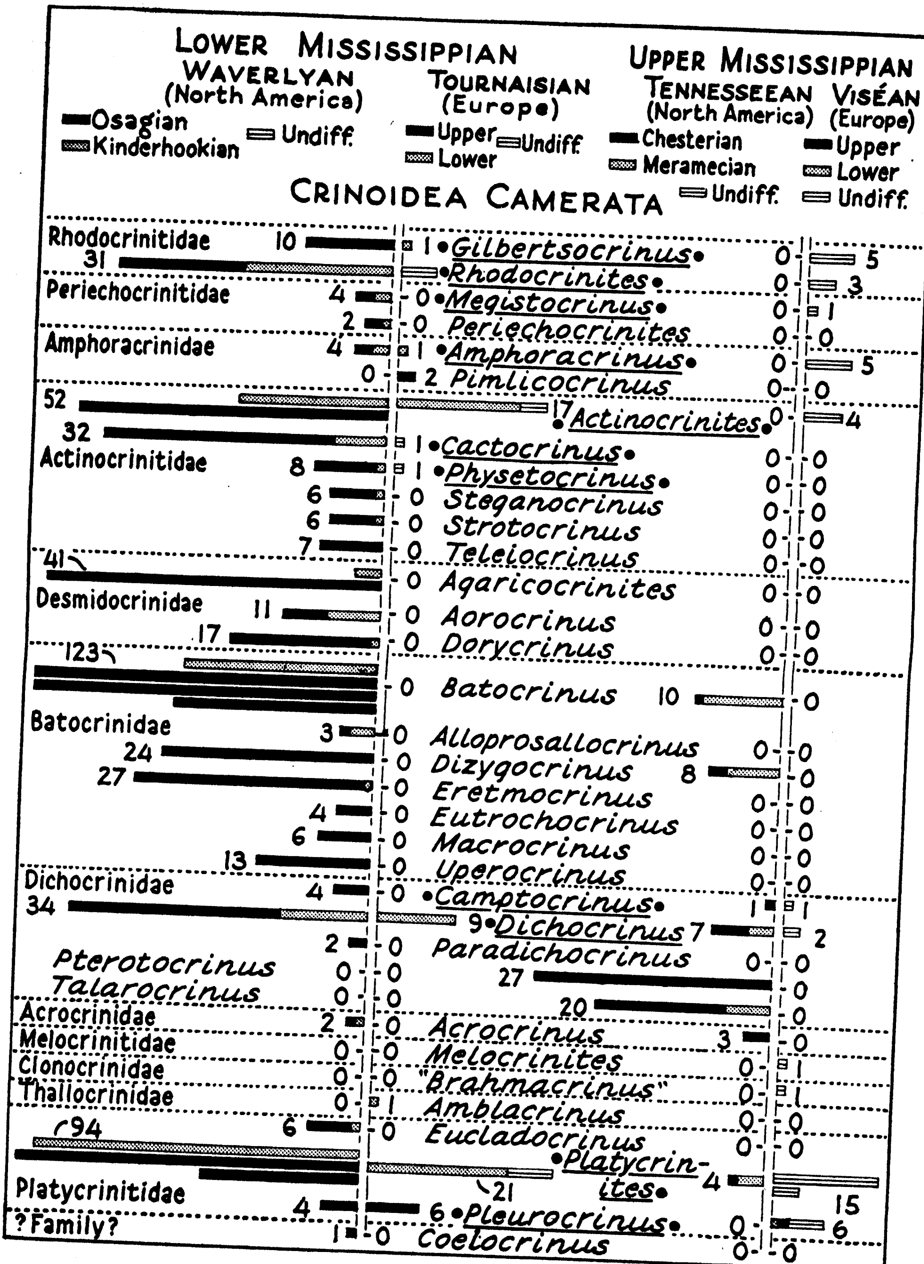



FIG. 12.—Comparison of Mississippian camerate crinoids from North America and Europe. Genera common to both continents are underlined. The disparity in distribution of the Desmidocrinidae, Batocrinidae, and part of the Dichocrinidae is especially noteworthy.

Upper Mississippian rocks as well. The fourth group is that of the platycrinitids, represented by the genera *Platycrinites* and *Pleurocrinus*. The record of this group is much like that of the dichocrinids. They are most abundant in Lower Mississippian rocks, but especially in Europe are well known in the Viséan formations. *Pleurocrinus* is recognized in the Permian of Timor.

The most remarkable contrasts in distribution of camerate crinoids are found in the families Desmidocrinidae, Batocrinidae, and Dichocrinidae. Although 69 species of desmidocrinids have been described from Lower Mississippian rocks of North America, not a single representative of this family is known in Europe. There are now some 218 described species of batocrinids, all of which are American, none European. They are assigned to 7 genera, but the greatly preponderant number (133) belongs to *Batocrinus*. Only 10 species of *Batocrinus* and 8 of *Dizygocrinus* are known to occur in rocks younger than Osagian. The record of the dichocrinids is partly quite different; among genera confined to North America, *Pterotocrinus* and *Talarocrinus*, which include 47 species, are exclusively Upper Mississippian, and *Paradichocrinus* is represented by two Osagian species. Evidently the prolific stocks included among desmidocrinids, batocrinids, and the *Talarocrinus-Pterotocrinus* part of the dichocrinids, originated in shallow seas of the North American interior and did not migrate into the European area.

### ECHINOIDS

Much less abundant than the crinoids in most Mississippian rocks but very interesting and important wherever they occur are remains of echinoids. My survey of these fossils indicates that at present 22 genera of Mississippian echinoids have been defined, of which 9 are common to North America and Europe (fig. 13). Excepting *Archaeocidaris*, several species of which have been described on the basis of individual plates and spines rather than on relatively complete tests, the fossils of this group are all of such morphologic complexity and differentiation that their occurrence stratigraphically must be deemed to have much significance. *Lepidechinus* among the lepidocentrids and *Pholidocidaris* among lepidesthids are found on both sides of the Atlantic in Lower Mississippian strata but have not been discovered in younger rocks. On the other hand, the genus *Melonechinus*, of the family Melonitidae, is an exclusively Upper Mississippian echinoid that is represented by 11 American species and 2 European forms. So far as known, the American species are from Meramecian formations. Other common genera—*Perischodomus*, *Lovenechinus*, *Palaechinus*, *Archaeocidaris*, and *Maccoya*—are recorded from both Lower and Upper Mississippian rocks. In general, stratigraphical distribution of the echinoids gives important confirming evidence to that derived from the camerate crinoids and some other echinoderms.

### BRYOZOANS

The bryozoans, called "polyzoans" in the British Isles, are generally similar in Mississippian rocks of North America and Europe, but this group of fossils does not now furnish much help in a comparative paleontological study. I have not undertaken to tabulate data on described genera and species. Much work remains to be done on American Mississippian Bryozoa, and probably the same is true in western Europe.

Fenestrellinids are reported in both Lower and Upper Mississippian rocks of

the British Isles and from Viséan strata on the Continent. Nekhoroshev (1932) has identified several species from Viséan rocks of western Germany with Osagian and Tennesseean species described from North America.

trasted with the absence of this genus in Lower Carboniferous rocks of western Europe. Also, *Sulcoretepora*, which is represented by several American species, seemingly is not recorded from Lower Carboniferous rocks of Europe, possibly

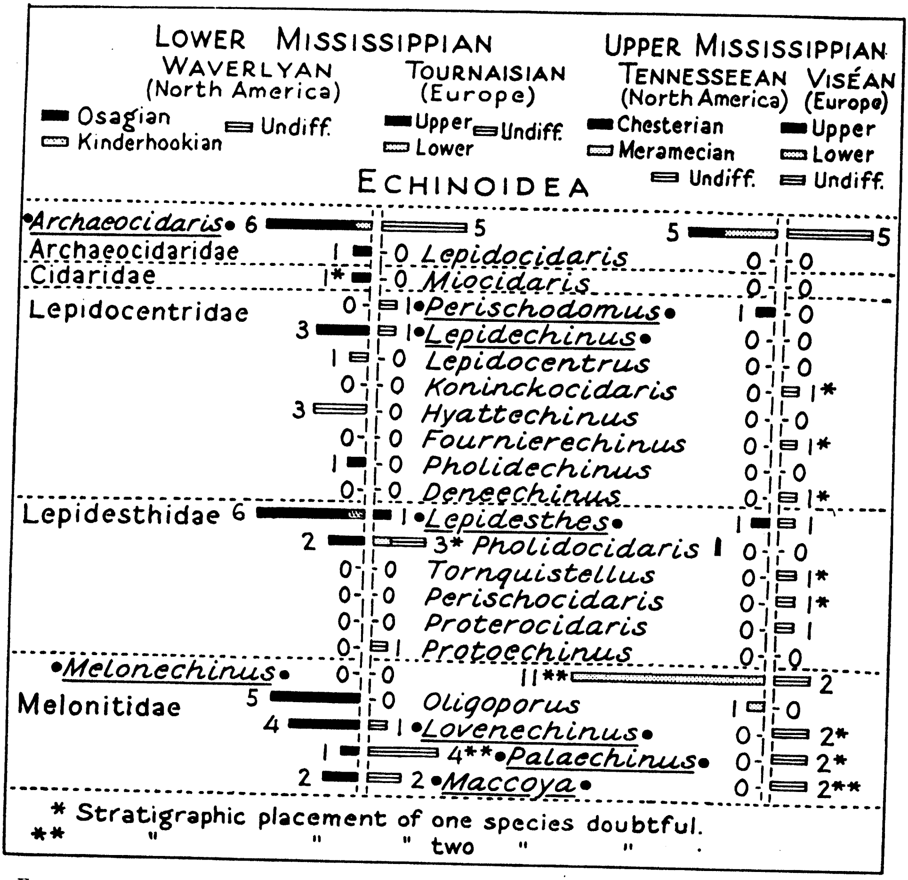



FIG. 13.—Comparison of Mississippian echinoids of North America and Europe. Genera common to both continents are underlined.

A noteworthy disparity in the distribution of bryozoans in rocks of Mississippian age on opposite sides of the Atlantic is the abundance of *Archimedes*, especially in Upper Osagian and Chesterian rocks of North America, as con-

excepting *Sulcoretepora* (?) *raricosta* McCoy from Lower Carboniferous rocks of Ireland. Other American bryozoan genera of distinctive type, unknown in Europe, are *Lyropora* and *Evactinopora*.

#### BRACHIOPODS

The brachiopods are a group having considerable value for the comparative study of the paleontology of Mississippian rocks in North America and Europe; but, because of only slight distinctions between many species that are defined, a useful report on this group cannot be made without much more study than I have been able to undertake. One may point out that, in general, the same broad stocks, including productids, chonetids, schuchertellids, schizophorids, spiriferids, syringothyrids, and rhynchonellids, are present in both successions and that there is also a general similarity of sequence. Among the productids, subgenera such as *Productus* (ss), *Buxtonia*, *Avonia*, *Echinoconchus*, *Linoproductus*, *Pustula*, *Productella*, and *Dictyoclostus* are recognized in Europe and North America. Also, several species assigned to these genera in the respective continents are closely similar. As indicated in figure 1, the productids and other brachiopods have been used in zoning the British Mississippian rocks, and they are recognized to be useful also in differentiating horizons of the Lower and Upper Mississippian in many parts of North America. Representative types of European Lower Carboniferous brachiopods are illustrated in figure 14.

The chonetids of western Europe include some forms very closely similar to *Chonetes logani* and *C. glenparkensis*, which are characterized by broad plications bearing cross-wrinkles; it is interesting to find some of these representatives of typical Kinderhookian species in America, along with *Leptaena analoga* and others, represented in lower Tournaisian rocks of Belgium and England. On the other hand, chonetids of similar type have been reported from Viséan strata of western Germany (Paeckelmann, 1930).

Differentiation of several Mississippian species of *Schizophoria*, as recorded in American paleontological literature and according to the work of some Europeans, is of very doubtful validity, and there is now a tendency to refer these various examples of *Schizophoria* to a single species.

Rhynchonellid genera, such as *Tetracamera*, *Shumardella*, *Paryphorhynchus*, and *Pugnoides*, which are widely distributed in Mississippian rocks of North America, seem not to have been recognized in western Europe.

Summarizing these observations somewhat roughly, one may say that in Tournaisian rocks there are many more similarities to Lower Mississippian rocks of North America than appear in the Viséan, where the appearance of *Gigantella*, *Daviesella*, and some other forms finds no equivalent on this side of the Atlantic, except for sporadic occurrence of *Gigantella*, associated with *Striatifera* and Viséan types of corals, along the Pacific border of this continent (Merriam, 1942; Sutton, 1938, p. 552). The noteworthy contrast between brachiopod assemblages of the Tournaisian and the Viséan and younger Mississippian strata of western Europe is hardly matched by contrasts between Osagian and Meramecian rocks of North America.

#### PELECYPODS AND GASTROPODS

The *Posidonia* or P zone of the upper Viséan in the British Isles is designated in terms of a common thin-shelled clam, which occurs mostly in dark shaly deposits. It is similar in form to *Caneyella*, which is found in beds of presumably equivalent age in the Caney and Moorefield shales of the central United States. I have collected *Posidonia* in association with ammonites in the Moorefield shale near Batesville, Arkansas.

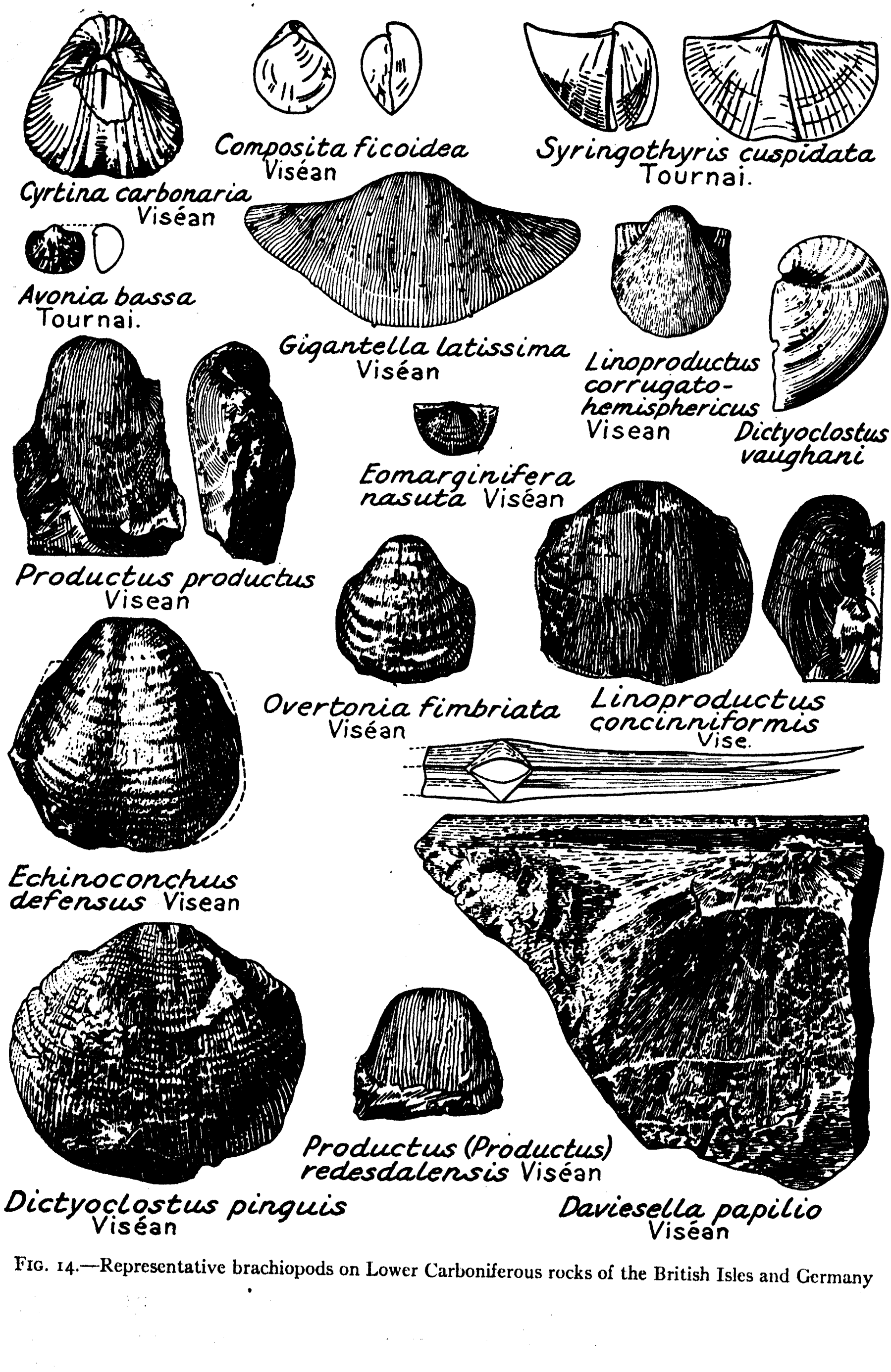



FIG. 14.—Representative brachiopods on Lower Carboniferous rocks of the British Isles and Germany

Another time-stratigraphic unit in the British area that is differentiated in terms of a pelecypod as guide fossil is the *Modiola lata* or Km subzone of the *Cleistopora* zone, of Etroeungtian age.

In general, neither pelecypods nor gastropods have yet been proved to have much importance for stratigraphic zonation or correlation of Mississippian deposits. Illustrative of this is extensive work by Hind (1896–1905) on a varied well-preserved molluscan fauna in the upper part of the Lower Carboniferous succession in Scotland; this was presumed to match closely the Nebraska City fauna described by Geinitz (1866) and Meek (1872), despite the fact that the Nebraska City fossils occur near the top of the Pennsylvanian succession in the northern Mid-Continent region.

Newell (1937) in studying the Pectinacea has noted the occurrence of two species of *Pterinopectinella* and one each of *Limipecten* and *Streblopteria* occurring in Lower Carboniferous strata of the British Isles; these shells are somewhat closely similar to forms known from the Mississippian of the United States. The genus *Obliquipecten*, which occurs in Viséan rocks of northern England, is not yet known outside the British area. Among myalinids (Newell, 1942) the genus *Myalina* ranges from Meramecian to Leonardian, *Septimyalina* from Chesterian and Viséan to Wolfcampian, and *Promytilus* from Viséan to Guadalupian. All three genera occur in Viséan rocks of England. A species of *Myalina* from the St. Louis limestone of the Mississippi Valley closely resembles *M. goldfussiana*, the genotype, from the Viséan of Europe. *Septimyalina angulata*, from Chester rocks of Illinois and Missouri, is rather closely similar to species assigned to this genus described by Hind (1896–1905) from England. Newell (1942, p. 52) has recognized a regular progression of morphologic features in the evolution of the mytilacean shells, indicating that these clams may have value in stratigraphic zonation.

Weller (1926) has defined the zone of *Sulcatopinna missouriensis* at the summit of Chesterian rocks in Illinois, and Laudon (1931) has differentiated Kinderhookian zones in terms of pelecypods and gastropods (*Palaeoneilo barrisi*, *Straparollus obtusus*, *Loxonema*) (fig. 2). According to present information, these zones are not known to have much value except locally.

### CEPHALOPODS

Cephalopods, especially ammonoids, which exhibit a wide variety of shell form, surface ornamentation, and suture pattern, are presumed to be among the most useful fossils for stratigraphic zonation and for correlation, even between basins on different continents. Extensive work in western Europe has demonstrated the value of these fossils, both for definition of a widely recognizable succession of zones and also in some areas for rather fine subdivision of these zones. The work of Bisat (1928, 1936) in England and of Schmidt (1928) and Delépine (1928) on the Continent is especially noteworthy and has made invaluable contributions to stratigraphic differentiation of the Culm or Kohlenkalk facies of the Viséan deposits and classification of the Namurian beds.

Inasmuch as the ammonoids are the subject of a special paper contributed to this symposium by A. K. Miller, I shall treat the group somewhat cursorily. Significant data of stratigraphic distribution and inferred phylogenetic relationships, based mainly on studies by Bisat, are presented in figures 15 and 16. In these diagrams species that are repre-

sented by identical or very closely similar American forms are indicated by an asterisk. Characteristic types of Lower Carboniferous ammonoids are illustrated in figure 17.

Observations that seem to me to have chief importance are the following:

1. The occurrence in western Europe and the United States of identical or very closely similar forms of *Protocanites*, *Imitoceras*, and *Muensteroceras*, found, respectively, in Tournaisian and Waverlyan rocks, clearly indicates a general time correspondence of these divisions.

2. The zone of *Beyrichoceras* is very clearly established in western Europe, both in the British Isles and on the Continent, as middle Viséan in age (figs. 1,

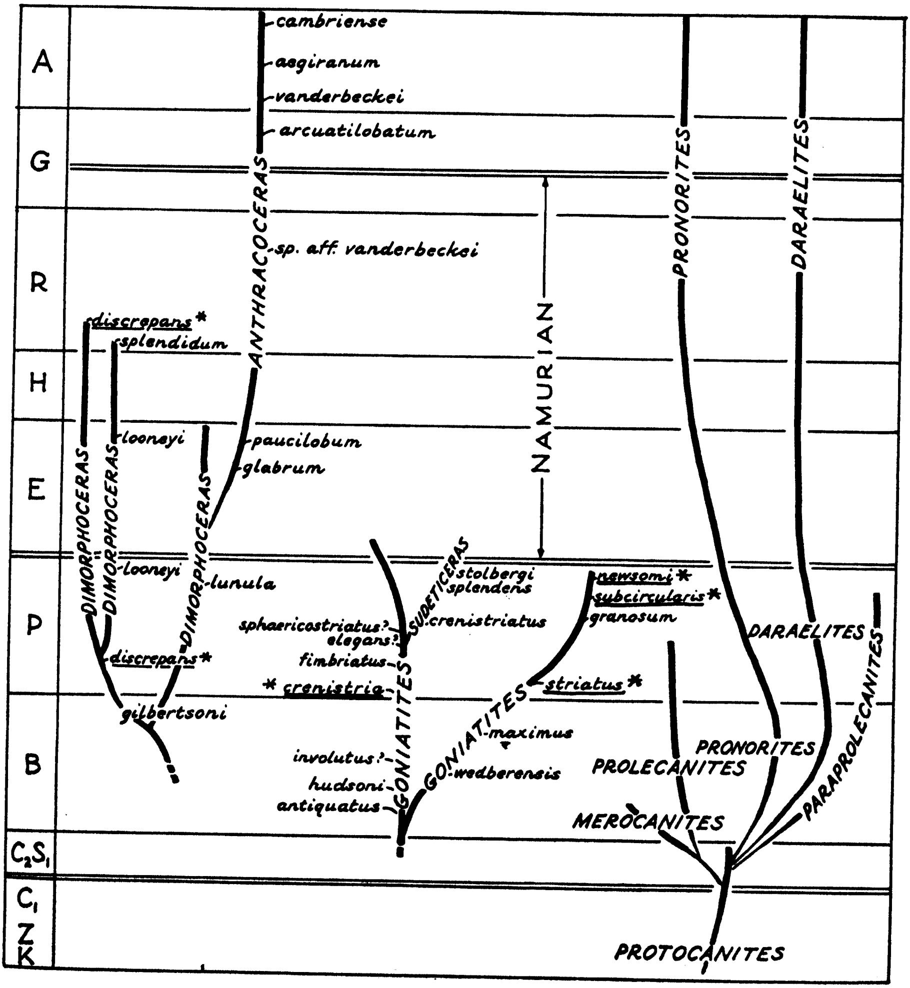



FIG. 15.—Viséan and Namurian ammonoid zones of the British Isles, showing distribution of species and inferred phylogeny of genera (data from Bisat).

15, 16). Beds belonging to this zone and containing its diagnostic fossils have been found to interfinger with strata of the coral-brachiopod zones in a manner to show that the $B_1$ subzone corresponds to the $S_2$ subzone containing numerous *Lithostrotion* and other characteristic Upper Mississippian corals, brachiopods, and other fossils (Hudson and Turner, 1933*a*); the $B_2$ subzone is found to correspond in age to the $D_1$ subzone, which very surely is not older than Meramecian deposits of North America and possibly belongs to the Lower Chesterian. Discovery of a representative of *Beyrichoceras* in the Osagian rocks of North America, as reported by Miller (1947), constitutes the record of a single identi-

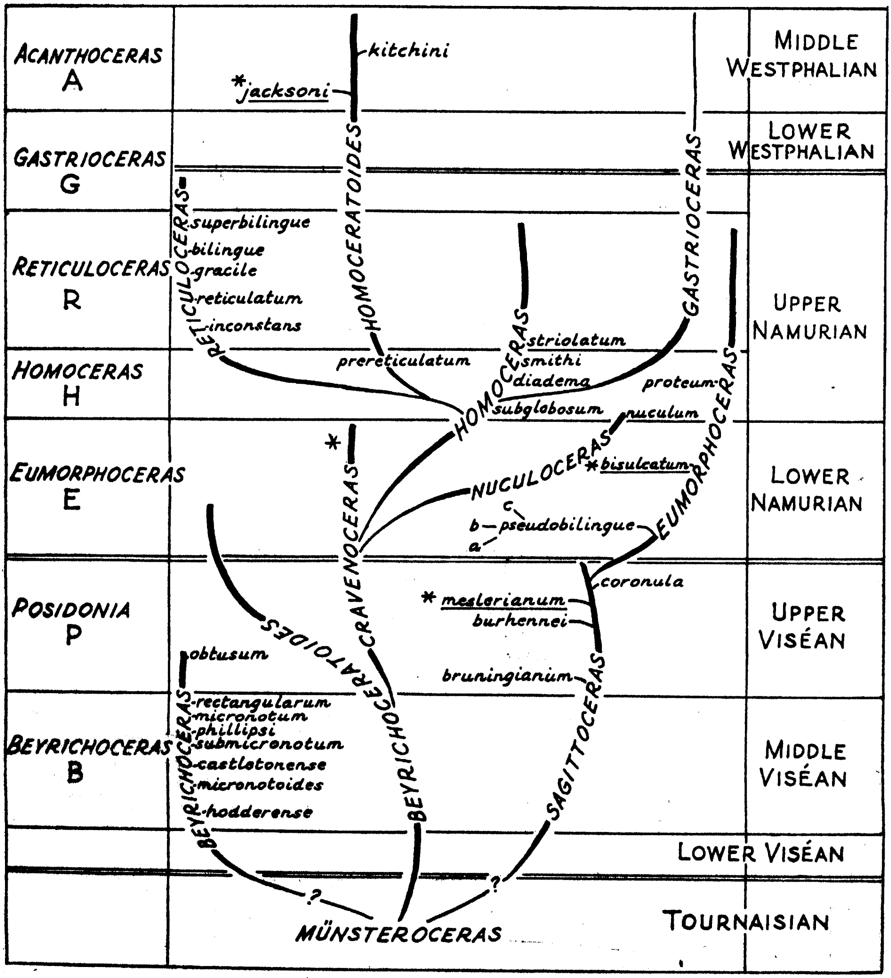



FIG. 16.—Viséan and Namurian ammonoid zones of the British Isles, showing distribution of species and inferred phylogeny of genera (data from Bisat).

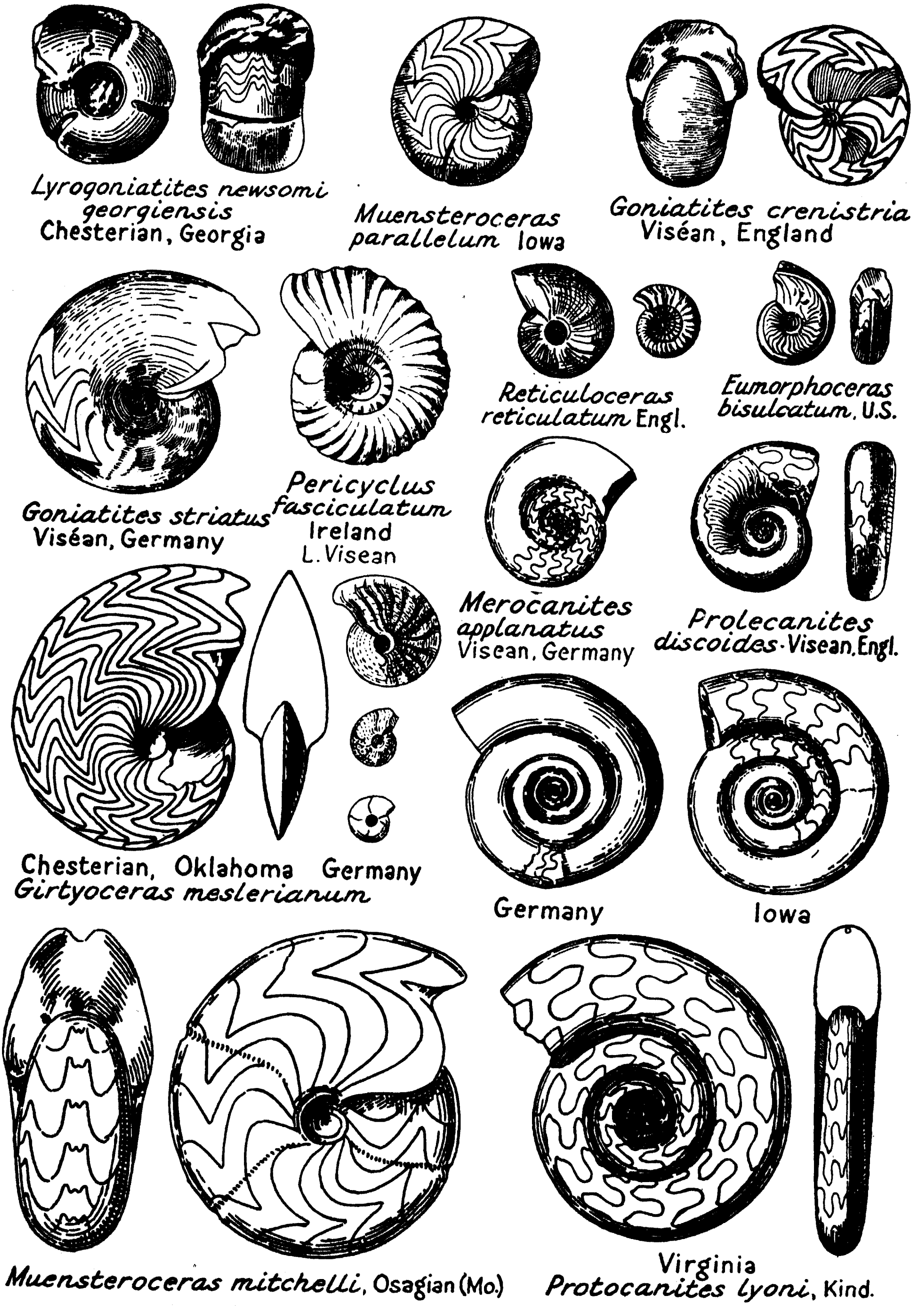



FIG. 17.—Representative Mississippian ammonoids from North America and western Europe (somewhat reduced).

fication of a representative of this genus in a stratigraphic position that can hardly fail to be far below that of the *Beyrichoceras* zone as defined in Europe. *Beyrichoceras* may have become differentiated as a genus in Tournaisian time, and the region in which it first became differentiated is unknown. As pointed out by Miller, the distinction between *Beyrichoceras* and *Muensteroceras* is extremely small, and the latter genus is very well developed in Tournaisian deposits. Overwhelming evidence supplied by crinoids, echinoids, brachiopods, corals, and probably other fossils, is utterly in conflict with the supposition that the European D zone can possibly be equivalent to pre-Meramecian deposits of North America. It is unfortunate that, as yet, numerous representatives of the European D zone have not been found on the western side of the Atlantic. Until such evidence is found, the record of *Beyrichoceras* in early Mississippian rocks of this continent cannot be permitted to outweigh abundant evidence of a differing sort.

3. Fossils of the *Posidonia* and *Eumorphoceras* zones in western Europe are distributed in well-defined successive arrangements that permit recognition of subzones. This seems not to be the case in North America, where characteristic P-zone fossils, such as *Goniatites crenistria*, are found in the selfsame layers as are diagnostic fossils of the E zone, such as *Eumorphoceras bisulcatum*. Rocks containing these fossils have been referred to upper Meramecian and to Chesterian parts of the Mississippian column, and the problem of their correct stratigraphic placement has not yet been solved.

## CONCLUSION

The paleontological comparison, somewhat cursorily made in this study, indicates many points of correspondence and some notable dissimilarities between the distributions of fossils in rocks of Mississippian age on opposite sides of the Atlantic. In the light of present knowledge the most important fossils for paleontological zonation and correlation are judged to be the corals, brachiopods, blastoids, crinoids, echinoids, and ammonoids. With varying exactitude these fossils have been employed for widely applicable zonation in western Europe; and the beginnings of an equally useful, widely applicable zonation in Mississippian rocks of North America have been made.

The comparative study strongly suggests that the most important line of partition in Mississippian rocks is that between the Tournaisian and Viséan, which on this continent corresponds to the line between Osagian and Meramecian strata. No satisfactory basis has yet been found for recognizing the position of the Kinderhookian-Osagian boundary in terms of the European succession. Likewise, placement of the Meramecian-Chesterian boundary in western Europe is in doubt; probably this belongs in the lower part of the D zone approximately corresponding to the boundary between the B and P zones of the ammonoid succession.

It is unsatisfactory that doubt must be recorded as to placement of the Mississippian-Pennsylvanian boundary in terms of the western European Carboniferous deposits. The lowermost Westphalian rocks assigned to the $G_2$ subzone, as indicated both by marine invertebrates and by land plants, is Early Pennsylvanian (Morrowan) in age. No equivalents of the *Homoceras* and *Reticuloceras* zones have been found in North America. It is possible that not only the upper part of the D zone but the

entire Namurian is of Chesterian age. In most sections, however, there is no pronounced break either stratigraphically or faunally between the Namurian and the overlying lower Westphalian. Locally, as in central England, there is noteworthy unconformity between the Carboniferous limestone series, with D beds at the top, and the succeeding Millstone grit of Namurian age; and accompanying this break is a marked change in lithology. On the other hand, Millstone grit sedimentation begins in some areas below the top of the D zone.

Intercontinental correlations between North America and Europe in the Lower Carboniferous part of the column are by no means definitely established.

## REFERENCES CITED


BASSLER, R. S. (1937) The Paleozoic rugose coral family Paleocyclidae: Jour. Paleontology, vol. 11, pp. 189–201, pls. 30–32.

——— and MOODEY, M. W. (1943) Bibliographic and faunal index of Paleozoic pelmatozoan echinoderms: Geol. Soc. America Special Paper 45, pp. 1–734.

BISAT, W. S. (1928) The Carboniferous goniatite zones of England and their continental equivalents: Cong. pour l'avancement des études de stratigraphie carbonifère, Heerlen, 1927, Compte rendu, pp. 117–133, pls. 6, 6A.

——— (1936) The faunal stratigraphy and goniatite phylogeny of the Carboniferous of western Europe, with notes on the connecting links with North America: Internat. Geol. Cong. Rept., 16th Session, Washington, 1933, pp. 529–537, tables 1–2.

COOPER, G. A. (1931) A new species of the echinoid *Lepidesthes*: Am. Jour. Sci., 5th ser., vol. 22, pp. 531–538, figs. 1–2.

DELÉPINE, G. (1928) Les Faunes du Dinantian de l'Europe occidentale: Cong. pour l'avancement des études de stratigraphie carbonifère, Heerlen, 1927, Compte rendu, pp. 223–233.

EASTON, W. H. (1943*a*) The fauna of the Pitkin formation of Arkansas: Jour. Paleontology, vol. 17, pp. 125–154, pls. 21–24, fig. 1.

——— (1943*b*) New Chester corals from Alabama and Tennessee: *ibid.*, vol. 17, pp. 276–280, pl. 46, fig. 1.

——— (1944) Corals from the Chouteau and related formations of the Mississippi Valley region: Illinois Geol. Survey Rept. Inv. 97, pp. 1–62, pls. 1–16, fig. 1.

——— (1945*a*) Kinkaid corals from Illinois: Jour. Paleontology, vol. 19, pp. 383–389, figs. 1–2.

——— (1945*b*) Corals from the Otter formation (Mississippian) of Montana: *ibid.*, pp. 522–528, figs. 1–10.

——— (1945*c*) Amplexoid corals from the Chester of Illinois and Arkansas: *ibid.*, pp. 625–632, pls. 85–87, fig. 1.

GEINITZ, H. B. (1866) Carbonformation und Dyas in Nebraska: Leopoldino-Carolinische deutsch. Akad. Naturf., Verk. 33, Abh. 4, pp. 1–91, plates.

HILL, DOROTHY (1938–1941) The Carboniferous rugose corals of Scotland: Palaeontographical Soc. Mon., pts. 1–2, pp. 1–78, pls. 1–2, 1938; pts. 3–5, pp. 79–114, pls. 3–5, 1939; pts. 6–11, pp. 115–204, pls. 6–11, 1940; pp. 205–213, 1941.

HIND, WHEELTON (1896–1905) Monograph of the British Carboniferous Lamellibranchiata: Palaeontographical Soc. Mon., vol. 1, 1896–1900; vol. 2, 1901–1905; pp. 1–708, plates.

HUDSON, R. G. S. (1942) An upper Viséan zaphrentoid fauna from the Yoredale beds of northwest Yorkshire: Yorkshire Geol. Soc. Proc., vol. 25, pp. 101–126, pls. 9–12.

——— (1944*a*) On the Carboniferous corals: *Zaphrentites shunnerensis*, sp. nov.: Geol. Mag., vol. 81, pp. 45–48, figs. 1–2.

——— (1944*b*) Lower Carboniferous corals of the genera *Rotiphyllum* and *Permia*: Jour. Paleontology, vol. 18, pp. 355–362, pls. 56–57, fig. 1.

——— (1945) On the Lower Carboniferous corals *Permia capax* and *P. rota*, n. sp.: Leeds Philos. Soc. Proc., vol. 4, pp. 285–298, pls. 1–2, figs. 1–2.

——— and TURNER, J. S. (1933*a*) Early and mid-Carboniferous earth movements in Great Britain: Leeds Philos. Soc. Proc., vol. 2, pp. 455–466, charts.

——— — ——— (1933*b*) Correlation of Dinantian and Namurian in western Europe: *ibid.*, vol. 2, pp. 467–482.

JEFFORDS, R. M. (1943) Caninia from the Lower Carboniferous of New Mexico: Jour. Paleontology, vol. 17, pp. 545–549, figs. 1–7.

KELLY, W. A. (1942) Lithostrotiontidae in the Rocky Mountains: Jour. Paleontology, vol. 16, pp. 351–361, pls. 50–51, fig. 1.

LAUDON, L. R. (1931) The stratigraphy of the Kinderhook series of Iowa: Iowa Geol. Survey Ann. Rept., vol. 35, pp. 333–452, figs. 45–68.

——— (1933) The stratigraphy and paleontology of the Gilmore City formation of Iowa: Iowa Univ. Studies, vol. 15, no. 2, pp. 1–74, pls. 1–7, figs. 1–6.

——— (1937) Stratigraphy of the northern extension of the Burlington limestone in Missouri and Iowa: Am. Assoc. Petroleum Geologists Bull. 21, pp. 1158–1167, figs. 1–4.

LEWIS, H. P. (1935) Lower Carboniferous corals of Nova Scotia: Annals and Mag. Nat. History, ser. 10, vol. 16, pp. 118–142, pls. 5–7, fig. 1.

LIEBUS, A. (1931) Die Fauna des deutschen Unterkarbons, 3 Teil: Die Foraminiferen: Preuss. geol. Landesanstalt Abh., neue Folge, no. 136, pp. 135–175, pls. 9–10.

MEEK, F. B. (1872) Report on the paleontology of eastern Nebraska: In HAYDEN, F. V., Final report of the U.S. Geol. Survey of Nebraska, 42d U.S. Cong., 1st Session, House Executive Doc. 19, pp. 83–239, plates.

MERRIAM, C. W. (1942) Carboniferous and Permian corals from central Oregon: Jour. Paleontology, vol. 16, pp. 372–381, pls. 54–57.

MILLER, A. K. (1947) A goniatite from the Mississippian Boone formation of Missouri: Jour. Paleontology, vol. 21, pp. 19–22, pl. 10, figs. 1–3.

———, and FURNISH, W. M. (1940) Studies of Carboniferous ammonoids, pts. 1–4: Jour. Paleontology, vol. 14, pp. 356–377, pls. 45–49.

MOORE, R. C., and LAUDON, L. R. (1943) Evolution and classification of Paleozoic crinoids: Geol. Soc. America Special Paper 46, pp. 1–153, pls. 1–14, figs. 1–18.

NEKHOROSHEV, B. (1932) Die Fauna des deutschen Unterkarbons, 3 Teil: Die Bryozoen des deutschen Unterkarbons: Preuss. geol. Landesanstalt Abh., neue Folge, no. 141, pp. 1–74, pls. 1–10, figs. 1–3.

NEWELL, N. D. (1937) Late Paleozoic pelecypods: Pectinacea: Kansas Geol. Survey, vol. 10, pt. 1, pp. 1–123, pls. 1–20, figs. 1–38.

——— (1942) Late Paleozoic pelecypods: Mytilacea: *ibid.*, pt. 2, pp. 1–115, pls. 2–115, figs. 1–22.

PAECKELMANN, W. (1930) Die Fauna des deutschen Unterkarbons, 2. Teil: Die Brachiopoden, 1. Teil: Die Orthiden, Strophomeniden und Choneten des mittleren und oberen Unterkarbons: Preuss. geol. Landesanstalt Abh., neue Folge, no. 122, pp. 143–326, pls. 9–24.

——— (1931) Die Fauna des deutschen Unterkarbons, 2 Teil: Die Productinae und Productus-ähnlichen Chonetinae: *ibid.*, no. 136, pp. 1–352, pls. 1–42.

SCHMIDT, HERMANN (1928) Biostratigraphie des Carbon in Deutschland: Cong. pour l'avencement des études de stratigraphie carbonifère, Heerlen, 1927, Compte rendu, pp. 663–672, pl. 16, fig. 1.

SLOSS, L. D. (1945) Corals from the post-Osage Mississippian of Montana: Jour. Paleontology, vol. 19, pp. 309–314, pl. 48.

SUTTON, A. H. (1938) Taxonomy of Mississippian Productidae. Jour. Paleontology, vol. 12, pp. 537–569, pls. 62–66, figs. 1–2.

WELLER, STUART (1926) Faunal zones in the standard Mississippian section: Jour. Geology, vol. 34, pp. 320–335.

# PUBLICATIONS RECEIVED

Annual Report of the Chicago Natural History Museum for 1946. Chicago, January, 1947.

Annual Report of the Director of the Department of Terrestrial Magnetism. By John A. Fleming. Reprinted from Carnegie Institution of Washington Year Book 45, for the Year, 1945–1946. Washington, 1946.

Annual Report 2, for the Fiscal Year July 1, 1946 to June 30, 1947, of the New Mexico Bureau of Mines and Mineral Resources. By E. C. Anderson. New Mexico School of Mines. Socorro, 1947.

Antimony Deposits of the Tejocotes Region of Oaxaca, Mexico. By Donald E. White and Reinaldo Guiza, Jr. U.S. Geological Survey Bulletin 953-A. Published in co-operation with Universidad Nacional Autonoma de Mexico, Instituto de Geologia, and Secretaria de la Economia Nacional Dirección General de Minas y Petroleo. Washington: U.S. Government Printing Office, 1947.

Barite, Fluorite, Galena, Sphalerite Veins of Middle Tennessee. By W. B. Jewell. State of Tennessee Department of Conservation, Division of Geology Bulletin 51. Nashville, 1947.

Bibliography of the Geology and Mineral Resources of Oregon (Supplement, July, 1936, to December, 1945). Compiled by John Eliot Allen. State of Oregon Department of Geology and Mineral Industries Bulletin 33. Portland, 1947.

Bibliography of the Geology and Water Resources of the Island of Hawaii. By G. A. Macdonald. Hawaii Division of Hydrography Bulletin 10. Honolulu, 1947.

Bibliography of Seismology. By Ernest A. Hodgson. Publications of the Dominion Observatory, vol. 13, no. 20, items 6143–6356. Ottawa, 1947.

Bibliography of Seismology. By W. G. Milne. Publications of the Dominion Observatory, vol. 14, no. 1, January to June, 1947. Ottawa, 1947.

Biography of the Earth: Its Past, Present, and Future. By George Gamow. Pelican Mentor Book. New York: New American Library, 1948.

Boletim da Sociedade Geologica de Portugal, vol. 6, no. 3. Porto, 1947.

British Regional Geology: The Central England District. By F. H. Edmunds and K. P. Oakley. Geological Survey and Museum, Department of Scientific and Industrial Research. 2d ed. London, 1947.

British Regional Geology: The Hampshire Basin and Adjoining Areas. By C. P. Chatwin. Geological Survey and Museum, Department of Scientific and Industrial Research. 2d ed. London, 1948.

British Regional Geology: London and Thames Valley. By R. L. Sherlock. Geological Survey and Museum, Department of Scientific and Industrial Research. 2d ed. London, 1947.

British Regional Geology: The South of Scotland. By J. Pringle. Geological Survey and Museum, Department of Scientific and Industrial Research. 2d ed., rev. Edinburgh, 1948.

British Regional Geology: The Welsh Borderland. By R. W. Pocock and T. H. Whitehead. Geological Survey and Museum, Department of Scientific and Industrial Research. 2d ed. London, 1948.

Bulletin of the Beach Erosion Board. War Department, Corps of Engineers, U.S. Army, vol. 1, no. 3; vol. 2, nos. 1, 2. Washington, D.C., 1947, 1948.

Bulletin of the Geological Society of China, vol. 24, nos. 3, 4. Published quarterly by the Society. Peiping: Ho Chi Press, May, 1946.

Calcita em franca Estado de São Paulo. By E. Fornasaro and M. Lancman. Summa Brasiliensis Geologiae, vol. 1, fasc. 3. 1946.

California Journal of Mines and Geology. State of California Department of Natural Resources, Division of Mines, vol. 43, no. 3. San Francisco, 1947.

California Mineral Production for 1946. By C. V. Averill, C. R. King, Henry H. Symons, and F. F. Davis. State of California Department of Natural Resources, Division of Mines Bulletin 139. San Francisco, 1948.

Carto geologica de la parte septentrional de la Republica Mexicana. By Philip B. King. Cartas Geologicas y Mineras de la Republica

Mexicana no. 3, in co-operation with the U.S. Geological Survey. Mexico, D.F., 1947.

The Causes and Effects of Sedimentation in Lake Decatur. By Carl B. Brown, J. B. Stall, and E. E. DeTurk. Illinois State Water Survey Division Bulletin 37. Urbana, 1947.

Cemented Sandstones of the Dakota and Kiowa Formations in Kansas. By Ada Swineford. University of Kansas Publications, State Geological Survey Bulletin 70, pt. 4. Lawrence, 1947.

Chromite Deposits near Red Lodge, Carbon County, Montana. By H. L. James. U.S. Geological Survey Bulletin 945-F. Washington: U.S. Government Printing Office, 1946.

The Classification and Origin of Plateaux, with Special Reference to India and the Adjacent Countries. By H. L. Chhibber. National Geographical Society of India Bulletin 8. Benares, 1948.

Composition and Properties of Petroleum in West Virginia. By A. J. W. Headlee, R. E. McClelland, E. J. Ball, and Helen H. Hess. State of West Virginia Geological and Economic Survey Report of Investigations 3-A. Morgantown, 1947.

Considerações sôbre a magnesita da serra das Eguas. By Reynaldo Saldanha da Gama. Summa Brasiliensis Geologiae, vol. 1, fasc. 2. 1946.

Contamination of Deep Water Wells in Southeastern Kansas. By Charles C. Williams. University of Kansas Publications, State Geological Survey Bulletin 76, pt. 2. Lawrence, 1948.

Contribução para o estudo de ***Hoplophorus euphractus*** Lund, 1839. By Carlos de Paula Couto. Summa Brasiliensis Geologiae, vol. 1, fasc. 4. 1947.

Deep Borings of Western South Dakota. By C. L. Baker. State Geological Survey Report of Investigations 57, University of South Dakota. Vermillion, 1947.

Desvaux Lake Area, Desserat Township, Rouyn-Noranda County. By P. E. Auger. Department of Mines, Mineral Deposits Branch Geological Report 27, Province of Quebec. Quebec: Redempti Paradis, 1947.

Some Deuteric Changes in the Enoggera Granite. By R. Gradwell. Proceedings of the Royal Society of Queensland, vol. 58, no. 4. Brisbane, 1947.

Dudley and Bridgnorth. By T. H. Whitehead and R. W. Pocock. Memoirs of the Geological Survey of Great Britain, Department of Scientific and Industrial Research. London, 1947.

Exploration for Oil and Gas in Western Kansas during 1946. By Walter A. Ver Wiebe. University of Kansas Publications, State Geological Survey Bulletin 68. Lawrence, 1947.

Field Tests for the Common Minerals. By George R. Fansett. Arizona Bureau of Mines, Technology series 42, Bulletin 154. 9th ed. Tucson: University of Ariziona, 1948.

The Geographical Names of Antarctica. Department of the Interior, U.S. Board on Geographical Names Special Publication 86. Washington, 1947.

Geologia e metalurgia. Publicação do Centro Moraes Rego, Bulletin 4. São Paulo: Praça Cel. Fernando Prestes, 74, 1946.

Geologic Features of the Connecticut Valley, Massachusetts, as Related to Recent Floods. By Richard H. Jahns. U.S. Geological Survey Water-Supply Paper 996. Prepared in co-operation with the Commonwealth of Massachusetts Department of Public Works. Washington: U.S. Government Printing Office, 1947.

Geologic Map of Ohio. By J. A. Bornocker. Geological Survey of Ohio. 3d printing. Scale, 1:500,000. Columbus, 1947.

Geological Explorations in the Island of Celebes under the Leadership of H. A. Brouwer. Amsterdam: North-Holland Publishing Company, 1947.

Geologie, part 1: Geologische Vorgänge der Gegenwart. By Herman Schmidt. Bücher der Mathematik und Naturwissenschaften, Notdruck. Wolfenbüttel-Hannover: Wolfenbütteler Verlagsanstalt G.m.b.H., 1947.

Geology and Artesian Water of the Alluvial Plain in Northwestern Mississippi. By Glen Francis Brown. Mississippi State Geological Survey Bulletin 65, in co-operation with the U.S. Geological Survey. University, 1947.

Geology and Geography of Karachi and Its Neighborhood. By Maneck B. Pithawalla and P. Martin-Kaye. Karachi, 1946.

Geology and Ground-Water Resources of Box Butte County, Nebraska. By R. C. Cady and O. J. Sherer. U.S. Geological Survey Water-Supply Paper 969. Prepared in co-operation with the Conservation and Survey Division, University of Nebraska. Washington: U.S. Government Printing Office, 1946.

Geology and Ground-Water Resources of Cedar City and Parowan Valleys, Iron County, Utah. By H. E. Thomas and G. H. Taylor.

U.S. Geological Survey Water-Supply Paper 993. Prepared in co-operation with the state of Utah. Washington: U.S. Government Printing Office. 1946.

Geology and Ground-Water Resources of the Island of Molokai, Hawaii. By H. T. Stearns and G. A. Macdonald. Hawaii Division of Hydrography Bulletin 11. Honolulu, 1947.

Geology and Ground-Water Resources of Kiowa County, Kansas. By Bruce F. Latta. University of Kansas Publications, State Geological Survey Bulletin 65. Lawrence, 1948.

Geology and Ground-Water Resources of Scott County, Kansas. By Herbert A. Waite. University of Kansas Publications, State Geological Survey Bulletin 66. Topeka, 1947.

Geology and Ground-Water Resources of Seward County, Kansas. By Frank E. Byrne and Thad G. McLaughlin. University of Kansas Publications, State Geological Survey Bulletin 69. Lawrence, 1948.

Geology and Mineral Deposits of the Baie Verte–Mings Bight Area. By Kenneth De Pencier Watson. Newfoundland Geological Survey Bulletin 21. St. John's, 1947.

Geology and Ore Deposits of Red River and Twining Districts, Taos County, New Mexico. By Charles F. Park, Jr., and Philip McKinley. New Mexico Bureau of Mines and Mineral Resources Circular 16. Prepared in co-operation with the U.S. Geological Survey. Socorro, 1948.

Geology in the Middle University District. By C. A. Cotton. From the University and the Community. 1946.

Geology of the Country around Weymouth, Swanage, Corfe, and Lulworth. By W. J. Arkell, with contributions by C. W. Wright and H. J. Osborne White. Geological Survey of Great Britain Memoirs, Department of Scientific and Industrial Research. London, 1947.

Geology of the Country around Witney. By Lindsdall Richardson, W. J. Arkell, H. G. Dines, with contributions by C. G. T. Morison. Geological Survey of Great Britain Memoirs, explanation of sheet 236. London, 1946.

Geology of the Dallas and Valsetz Quadrangles, Oregon. By Ewart M. Baldwin. State of Oregon, Department of Geology and Mineral Industries Bulletin 35. Portland, 1947.

The Geology of the Flora Quadrangle. By John R. Branch. North Dakota Geological Survey Bulletin 22. Grand Forks, 1947.

Geology of the Green River Desert–Cataract Canyon Region, Emery, Wayne, and Garfield Counties, Utah. By Arthur A. Baker. U.S. Geological Survey Bulletin 951. Washington: U.S. Government Printing Office, 1946.

Geology of the Highwood-Elbow Area, Alberta. By J. A. Allan and J. L. Carr, with appendix on paleontology by P. S. Warren. Research Council of Alberta Report 49, Province of Alberta. Edmonton: A. Shnitka, 1947.

Geology of the Lizard and Meneage. By J. S. Flett. Geological Survey of Great Britain Memoirs, explanation of sheet 359. London, 1946.

Geology of a Middle Devonian Cannel Coal from Spitsbergen. By Thorolf Vogt. Reprinted from Norsk geologisk tidsskrift, vol. 21. Oslo: I Kommisjon hos Jacob Dybwad, 1941.

Geology of Reef Ridge, Coalinga District, California. By Ralph Stewart. U.S. Geological Survey Professional Paper 205-C. Washington: U.S. Government Printing Office, 1946.

Glacial Studies of the Pleistocene of North America. By William H. Hobbs. Ann Arbor: J. W. Edwards, 1947.

Graphic Representation of Oil-Field Brines in Kansas. By Russell M. Jeffords. University of Kansas Publications, State Geological Survey Bulletin 76, pt. 1. Lawrence, 1948.

The Green World of the Naturalists. By Victor Wolfgang von Hagen. New York: Greenberg, Publisher, 1948.

Ground-Water Conditions in the Monroe Area, Louisiana. By P. H. Jones and C. N. Holmes. Department of Conservation, Louisiana Geological Survey Geological Bulletin 24. Baton Rouge, 1947.

Hidrogeologia y minerales no-metalicos de la zona norte del Estado de Michoacán. By Luis Blasquez L. and Raul Lozano Garcia. Anales del Instituto de Geologia, vol. 9. Mexico, D. F., 1946.

A Short History of the Tasman Geosyncline of Eastern Australia. By W. R. Browne. University of Sydney, Geology Department Publication 84, new series. Reprinted from Science Progress, vol. 35. London, 1947.

Investigations on Secondary Recovery by the Illinois State Geological Survey. By Alfred H. Bell. Illinois State Geological Survey Circular 129. Urbana, 1947.

Itawamba County Mineral Resources. By Franklin Earl Vestal and Harry J. Knollman.

Mississippi State Geological Survey Bulletin 64. University, 1947.

Kansas Clay, Dakota Formation. By Norman Plummer and John F. Romary. University of Kansas Publications, State Geological Survey Bulletin 67. Lawrence, 1947.

The Late Pleistocene Loesses of Central Kansas. By John C. Frye and O. S. Fent. University of Kansas Publications, State Geological Survey Bulletin 71, pt. 3. Lawrence, 1947.

Local Floods in Ohio during 1947. By William P. Cross. Ohio Water Resources Board Bulletin 14, Prepared in co-operation with the U.S. Geological Survey. Columbus, 1948.

On *Lufengosaurus magnus* Young (sp. nov.) and Additional Finds of *Lufengosaurus huenei* Young. By C. C. Young. Palaeontologia Sinica, National Geological Survey of China, new series C, no. 12; whole series 132. Nanking, 1947.

Magnetic Survey of Southeastern Crawford County, Kansas. By Robert M. Dryer. University of Kansas Publications, State Geological Survey Bulletin 70, pt. 5. Lawrence, 1947.

Major Tectonic Phenomena and the Hypothesis of Convection Currents in the Earth. By F. A. Vening Meisesz. Third William Smith Lecture. Reprinted from the Quarterly Journal of the Geological Society of London, vol. 103. London, 1948.

Major Winter and Nonwinter Floods in Selected Basins in New York and Pennsylvania. By Walter B. Langbein and others. U.S. Biological Survey Water-Supply Paper 915. Prepared in co-operation with the Federal Emergency Administration of Public Works. Washington: U.S. Government Printing Office, 1947.

Manganese Deposits of the Republic of Haiti. By E. N. Goddard, L. S. Gardner, and W. S. Burbank. U.S. Geological Survey Bulletin 953-B. Washington: U.S. Government Printing Office, 1947.

Manner of Locating and Holding Mineral Claims in California. By A. H. Ricketts, revised by C. A. Logan. State of California Department of Natural Resources, Division of Mines. Preprint from California Journal of Mines and Geology, vol. 44, no. 1. January, 1948.

Mineral Industry of California in 1947. State of California Department of Natural Resources, Division of Mines. Preprint from the California Journal of Mines and Geology, vol. 43, no. 4. October, 1947.

Minerals of Arizona. By Frederic W. Galbraith. Arizona Bureau of Mines, Geological Series 17, Bulletin 153. Tucson: University of Arizona, 1947.

Mines and Prospects of the Mount Reuben Mining District, Josephine County, Oregon. By Elton A. Youngberg. State of Oregon Department of Geology and Mineral Industries Bulletin 34. Portland, 1947.

The Mining Industry of the Province of Quebec in 1945. Department of Mines, Province of Quebec. Quebec: Redempti Paradis, 1946.

Mining Review for the Half-Year Ended 30th June, 1946, no. 84. Issued under the authority of the Hon. A. Lyell McEwin, Minister of Mines, South Australia Department of Mines. Adelaide: K. M. Stevenson, 1947.

Mining Review for the Half-Year Ended 31st December, 1946, no. 85. Issued under the authority of the Hon. A. Lyell McEwin, South Australia Department of Mines. Adelaide: K. M. Stevenson, 1947.

Miscelánea almera. Diputación Provincial de Barcelona, Publicaciones del Instituto Geológico 7. Barcelona, 1945.

New England's Buried Treasure. By Clay Perry. New York: Stephen Daye Press, 1946.

New Mexico Oil and Gas Production Data for 1946 (exclusive of Lee County). By N. Raymond Lamb and W. B. Macy. New Mexico Bureau of Mines and Mineral Resources and New Mexico Oil Conservation Commission Circular 16. Socorro, 1947.

Novo habito de berilo em jazida brasileira. By Reynaldo Saldanha da Gama. Summa Brasiliensis Geologiae, vol. 1, fasc. 1. 1946.

The Occurrence, Composition, Testing, and Utilization of Underground Water in South Australia, and the Search for Further Supplies. By L. Keith Ward. Department of Mines, Geological Survey of South Australia Bulletin 23. Adelaide: K. M. Stevenson, 1946.

Oil and Gas Development in Illinois in 1946. By Alfred H. Bell and Virginia Kline. Illinois State Geological Survey, Press Bulletin, series 56. Urbana, 1947.

Oil and Gas Map of Louisiana. Compiled by G. O. Coignet. Compiled by Louisiana Geological Survey. Published by Department of Conservation. October, 1947.

Ore Genesis of Queensland. By O. A. Jones. Proceedings of the Royal Society of Queensland, vol. 59, no. 1. Brisbane, 1947.

Origem das rochas alcalinas. By Djalma Guimarães. Estado de Minas Gerais, Instituto de Technologia Industrial Bulletin 5. Belo Horizonte: Grafica Queiroz Breiner Ltda., 1947.

Peat in Quebec: Its Origin, Distribution, and Utilization. By H. Girard. Department of Mines, Mineral Deposits Branch, Geological Report 31. Quebec: Redempti Paradis, 1946.

Petrology of a Middle Devonian Cannel Coal from Spitsbergen. By Gunnar Horn. Reprinted from Norsk geologisk tidsskrift, vol. 21. Oslo: I Kommisjon hos Jacob Dybwad, 1941.

Preliminary Note on the Nature of the Stresses Involved in the Late Paleozoic Diastrophism in New South Wales. By S. W. Carey and G. D. Osborne. University of Sydney, Geology Department Publication 31, new series. Reprinted from the Journal and Proceedings of the Royal Society of New South Wales, vol. 72. Sydney, 1939.

Proceedings of the Geological Society of South Africa, January to December, 1944. Edited by the Honorary Secretary. Johannesburg: Hortors Ltd., 1945.

Publications on the Geology and Mineral Resources of Virginia. By Arthur Bevan. Virginia Conservation Commission, Virginia Geological Survey Circular 6. University, Va., 1947.

Pumice Aggregate in New Mexico: Its Uses and Potentialities. By Donn M. Clippinger and Walter E. Gay. New Mexico Bureau of Mines and Mineral Resources Bulletin 28. Socorro, 1947.

Report of the Geological Survey Board for the Year 1945, with a Note on the Work of the Geological Survey and Museum during the Years 1939–1944. Department of Scientific and Industrial Research. London, 1947.

Report of the State Geologist on the Geology and Mineral Industries of Vermont, 1945–1946. By Elbridge C. Jacobs. Centennial issue. Geological Survey of Vermont. Burlington, 1946.

Reservoir Sedimentation in the Sacramento–San Joaquin Drainage Basins, California. By Carl B. Brown and Eldon M. Thorp. U.S. Department of Agriculture, Soil Conservation Service, Special Report 10. Washington, 1947.

Review of the Carboniferous Stratigraphy, Tectonics, and Paleogeography of New South Wales and Queensland. By S. W. Carey and W. R. Browne. University of Sydney, Geology Department Publication 21, new series. Reprinted from Journal and Proceedings of the Royal Society of New South Wales, vol. 71. Sydney, 1938.

Review of Petroleum Geology in 1946. By F. M. Van Tuyl and W. S. Levings. Quarterly of the Colorado School of Mines, vol. 42, no. 3. Golden, 1947.

Sea Islands of Georgia. By Count D. Gibson. Athens: University of Georgia Press, 1948.

Siscoe Mine Map-Area, Dubuisson and Vassan Townships, Abitibi-East County. By P. E. Auger. Department of Mines, Mineral Deposits Branch Geological Report 17, Province of Quebec. Quebec: Redempti Paradis, 1947.

Soil Conservation. Pennsylvania Department of Commerce State Planning Board, vol. 10, no. 2. Harrisburg, 1947.

A Stone Age Cave Site in Tangier: Preliminary Report on the Excavations at the Mugharet el Aliya, or High Cave, in Tangier. By Bruce Howe and Hallam L. Movius, Jr. Papers of the Peabody Museum of American Archaeology and Ethnology, Harvard University, vol. 28, no. 1. Cambridge, 1947.

The Stratigraphy and Structure of the Silurian and Devonian Rocks of Yass-Bowning District, New South Wales. By Ida A. Brown. University of Sydney, Geology Department Publication 42, new series. Reprinted from the Journal and Proceedings of the Royal Society of New South Wales, vol. 74. Sydney, 1941.

Subsurface Geologic Cross Section from Baca County to Yuma County, Colorado. By John C. Maher. University of Kansas Publications, State Geological Survey Oil and Gas Investigations Preliminary Cross Section 6. Lawrence, 1948.

Subsurface Geologic Cross Section from Ford County, Kansas, to Dallam County, Texas. By Fanny Carter Edson. University of Kansas Publications, State Geological Survey Oil and Gas Investigations Preliminary Cross Section 3. Lawrence, 1947.

Subsurface Geologic Cross Section from Scott County, Kansas to Otero County, Colorado.

By John C. Maher. University of Kansas Publications, State Geological Survey Oil and Gas Investigations Preliminary Cross Section 4. Lawrence, 1947.

Subsurface Geologic Cross Section from Trego County, Kansas, to Cheyenne County, Colorado. By Jack B. Collins. University of Kansas Publications, State Geological Survey Oil and Gas Investigations Preliminary Cross Section 5. Lawrence, 1947.

Surface Water Supply of the United States, 1945. Part 7: Lower Mississippi River Basin. U.S. Geological Survey Water-Supply Paper 1037. Prepared in co-operation with the states of Arkansas, Colorado, Kansas, Kentucky, Louisiana, Mississippi, Missouri, New Mexico, Oklahoma, Tennessee, and Texas and other agencies. Washington: U.S. Government Printing Office, 1947.

Surface Water Supply of the United States, 1945. Part 8: Western Gulf of Mexico Basins. U.S. Geological Survey Water-Supply Paper 1038. Prepared in co-operation with the states of Colorado, Louisiana, New Mexico, and Texas and other agencies. Washington: U.S. Government Printing Office, 1947.

Surface Water Supply of the United States, 1945. Part 9: Colorado River Basin. U.S. Geological Survey Water-Supply Paper 1039. Prepared in co-operation with the states of Arizona, Colorado, New Mexico, Utah, and Wyoming and other agencies. Washington: U.S. Government Printing Office, 1947.

Surface Water Supply of the United States, 1945. Part 11: Pacific Slope Basins in California. U.S. Geological Survey Water-Supply Paper 1041. Prepared in co-operation with the states of California and Oregon and other agencies. Washington: U.S. Government Printing Office, 1947.

Surface Water Supply of the United States, 1945. Part 14: Pacific Slope Basins in Oregon and Lower Columbia River Basin. U.S. Geological Survey Water-Supply Paper 1044. Prepared in co-operation with the states of Oregon and Washington and other agencies. Washington: U.S. Government Printing Office, 1947.

Talc Deposits of Murray County, Georgia. By A. S. Furcron, Kefton H. Teague, and James L. Calver. Georgia State Division of Conservation, Department of Mines, Mining and Geology, Geological Survey Bulletin 53. Atlanta, 1947.

Tonnancourt-Holmes Map-Area. Abitibi County. By W. W. Longley. Department of Mines, Division of Geological Surveys, Geological Report 24. Quebec: Redempti Paradis, 1946.

Tungsten Deposits of Vance County, North Carolina, and Mecklenburg County, Virginia. By G. H. Espenshade. U.S. Geological Survey Bulletin 948-A. Washington: U.S. Government Printing Office, 1947.

Utilization of Surface Water Resources of Sevier Lake Basin, Utah. By Ralf R. Woolley. U.S. Geological Survey Water-Supply Paper 920. Washington: U.S. Government Printing Office, 1947.

La Valeur de l'interprétation géologique de profils magnétiques. By Ivan de Magnée and Pierre Evrard. Société géologique Belgique Annales, vol. 69. Liége: H. Vaillant-Carmanne, S.A., 1945.

Vanadium Deposits near Placerville, San Miguel County, Colorado. By. R. P. Fischer, J. C. Haff, and J. F. Rominger. Colorado Scientific Society Proceedings, vol. 15, no. 3. Denver, 1947.

Water Levels and Artesian Pressure in Observation Wells in the United States in 1944. Part 2: Southeastern States. By A. N. Sayre and others. U.S. Geological Survey Water-Supply Paper 1017. Prepared in co-operation with the states of Alabama, Florida, Georgia, Kentucky, Maryland, Mississippi, North Carolina, Tennessee, Virginia, and West Virginia and other agencies. Washington: U.S. Government Printing Office, 1947.

The Water Resources of Tuscarawas County, Ohio. By James W. Cummins and Earl E. Sanderson. Ohio Water Resources Board Bulletin 6. Columbus, 1947.

VOLUME 56 NUMBER 5

THE JOURNAL OF GEOLOGY

*September 1948*

# ANCIENT ARCTICA[1]

A. J. EARDLEY
University of Michigan

ABSTRACT

"Ancient Arctica" is the name here applied to the lands and seas of the past in the Arctic region. A survey was made of the shields that face and partly surround the Arctic Sea, of the intra-shield orogenic belts that extend into and under the sea, of the troughs of deposition and orogenic belts of Alaska, and of the topography of the Arctic Sea floor. It is concluded that the region is underlain by continental crustal material chiefly of shield character; that appreciable changes have occurred in the distribution of land and water by epeirogenic and orogenic movements of the crust; that the present deep basin of the Arctic Sea began to sink in Carboniferous time; that the Alaska–northeastern Siberia region was one of nearly constant land connection during Mesozoic and Cenozoic time and probably also during the Paleozoic; and that the crust of the North Atlantic and Greenland Sea underwent movements in Tertiary time sufficient to provide land bridges for migration of animals and plants between Europe and North America, possibly on several occasions.

Brief reference is made to the problems of biogeography, and the bearing that the theory of Ancient Arctica has on some of these problems.

The lands of the Arctic hold good petroleum possibilities, and those of the Arctic archipelago are singled out for special discussion because of their particular promise.

## INTRODUCTION

### DEFINITION OF ARCTICA

The word "Arctica" is here defined to include (1) Alaska, (2) the Arctic archipelago north of the North American continent, (3) the great island of Greenland, (4) the shelf areas adjacent to the continents that border the Arctic Sea, (5) the deep basin of the Arctic Sea, (6) the shield areas of the various continents that face the Arctic Sea, and (7) the orogenic belts that impinge on the Arctic Sea or extend into and under it (fig. 1). A review of the geology of the lands of Arctica and of the topography of the sea floor points to the existence from time to time of land in places where the waters now spread. To the lands and seas of the north polar region of the geologic past the name "Ancient Arctica" is given.

### PURPOSE OF PAPER

The purpose of this paper is to review the geology of Arctica in order to pursue the problem of the relation of Eurasia to North America and Greenland. The conclusions will have an obvious bearing on the problems of biogeography, both past and present. They may also have an important bearing on exploration for oil and gas in the Arctic.

## ALASKA

### PALEOZOIC GEOSYNCLINE AND RELATED OROGENY

Most of the Paleozoic rocks of Alaska are exposed in the Brooks Range, Seward Peninsula, the central Yukon drainage

[1] Manuscript received April 20, 1948.

area (Central Plateau), and the Alexander Archipelago (fig. 2). The Alaska, Nutzotin, and Wrangell ranges also contain Paleozoic rocks, and, near by, a belt extends along part of Copper River and Chitina River valleys. The towering mass of Mount McKinley in the Alaska Range is eroded mostly from deformed Paleozoic strata.

For a detailed study of the Paleozoic rocks of Alaska, Smith's U.S. Geological Survey Professional Paper 192 should be consulted, particularly the large correlation chart in the pocket. The formations are well exposed in the Central Plateau, and a résumé of a few selected sections is given in table 1.

The igneous rocks in the Yukon-Tanana region have been summarized by Mertie (1935, p. 302) as follows: Basic lavas of basaltic and diabasic character were extruded during at least five geologic epochs in the Paleozoic. The first was in the Middle Ordovician, the second in the Middle Devonian, and the last three during three epochs of the Carboniferous. Granular intrusives of the same general character accompanied the extrusion of the lavas, but the volume of such rocks is relatively small. Some rhyolite and dacite lavas and tuffs are found among the Carboniferous lavas, but, in general, lavas of acidic or intermediate character are rare. Ultrabasic rocks were intruded during the Upper Devonian epoch.

The volcanism that, according to Mertie, occurred during the Carboniferous period in Alaska was greater than in any other period and most intense in the Alaska Range. Eruptions of basic lavas accompanied epeirogenic movements that persisted into the Triassic.

It is immediately clear that the rocks of part of the Central Plateau and of the southern part of Alaska represent the volcanic archipelago assemblage previously recognized and described in the western Cordillera of southeastern Alaska, British Columbia, Washington, Oregon, California, and Nevada (Kay, 1947, p. 1290; and Eardley, 1947, p. 342). The presence of the basic intrusives in the volcanic assemblage suggests that the belt was the site of both intrusive and extrusive activity and not simply a trough adjacent to a volcanic archipelago.

In northern Alaska the Paleozoic rocks are mainly sandstones, shales, and limestones and are typical of the mainland assemblage, also previously described in the western Cordillera. No volcanic rocks have been found in the sediments.

If a line is drawn separating the volcanic archipelago assemblage from the mainland assemblage, it will appear somewhat like that of figure 1. The curvature of the present mountain systems has influenced the curvature given by the writer to the line, but, then, there is considerable value to the premise that the belts of deformation trend approximately parallel with the former troughs of sedimentation.

The distribution in outcrop and the geosynclinal thicknesses, where known, of the Paleozoic strata indicate that the whole of Alaska was a region of subsidence and sedimentation in the Paleozoic as an extension of the Cordilleran geosyncline. Where did the sediments of the mainland assemblage come from? The only possible source must have lain to the north where the Arctic Sea now spreads.

### TRIASSIC AND JURASSIC GEANTICLINE AND ADJACENT BASINS

During the Triassic period and persisting into the Jurassic a great geanticline rose from the Paleozoic geosyncline and separated two adjacent basins of ac-

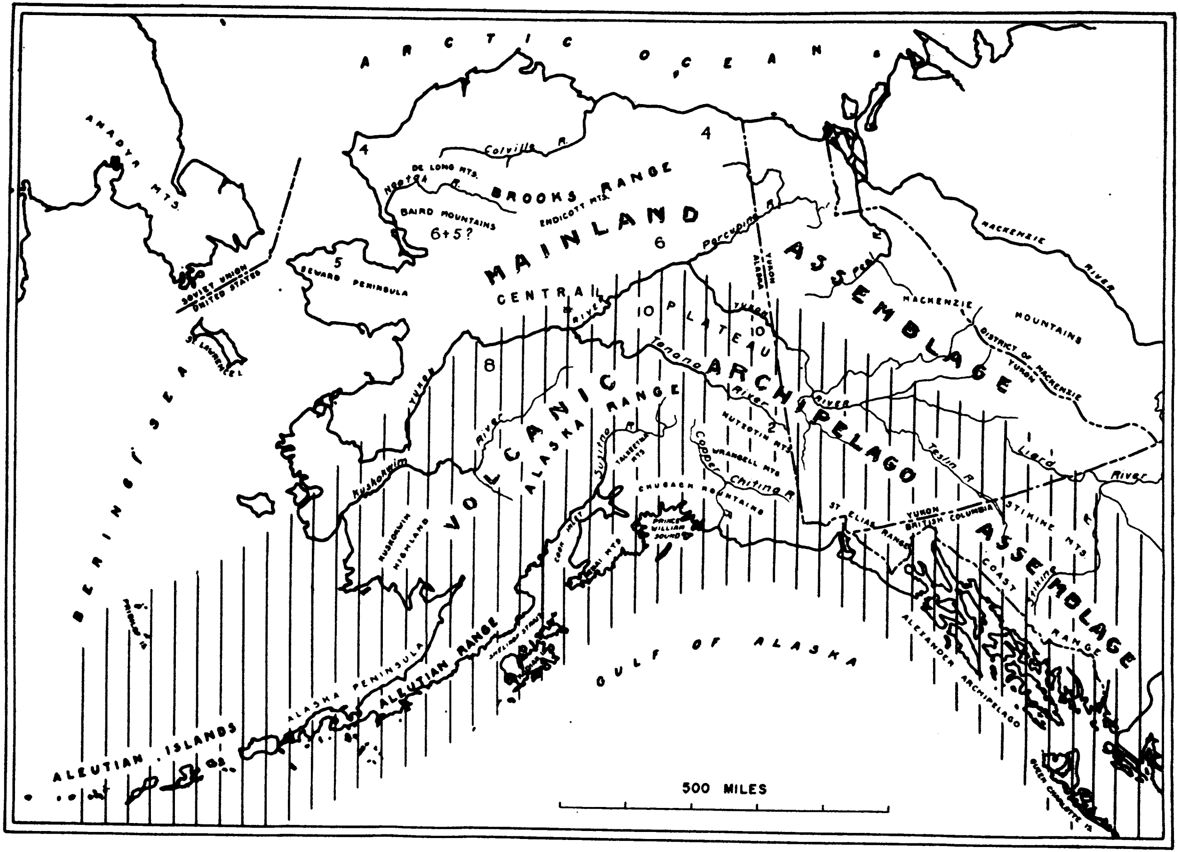



FIG. 2.—Alaskan Paleozoic geosynclinal divisions. Lined area is the volcanic archipelago assemblage, and the blank area is the mainland assemblage. Numbers represent thousands of feet of sediments and volcanics in those few places where estimates have been made. Generally, they represent only part of the Paleozoic section.

## TABLE 1

SELECTED PALEOZOIC SECTIONS OF ALASKA*

| Section | Description | Thickness |
|---|---|---|
| | **Permian** | |
| Kandik district | Limestone, conglomerate, shale and sandstone; Tahkandit limestone | .................... |
| | **Mississippian** | |
| Porcupine district | Dark shale and limestone, in part same as Calico Bluff formation | .................... |
| Koyukuk-Melozi district | Greenstone, little rhyolite, formerly considered part of Kanuti group | .................... |
| Yukon-Tanana district | Clay shale, sandstone, conglomerate; Nation River formation | 4,000– 6,000 feet |
| | Lava flows and associated sediments; Rampart group and Circle volcanics | 5,000–10,000 feet |
| | Limestone, shale, slate; Calico Bluff formation<br>Limestone beds<br>Undifferentiated schist, shale, chert, quartzite | 13,000 feet |
| | Chert with minor amounts of limestone and shale; Livengood chert | 2,000– 4,000 feet |
| | **Upper Devonian** | |
| Eagle district | Basalt lava and pyroclastics of greenstone habit; Woodchopper volcanics | 10,000± feet |
| | **Middle Devonian** | |
| Kantishna Nenana district | Massive limestone, equivalent to part of Tonzona | .................... |
| | Limy shale, more calcareous at top | .................... |
| | Shale, argillite, graywacke, quartzite | 10,000± feet |
| | Black conglomerate, white conglomerate, shale, and graywacke | .................... |
| | **Middle Silurian** | |
| Porcupine district | Black fissile shale, little siliceous limestone<br>Buff magnesian limestone | 2,500+ feet |
| | **Middle Ordovician** | |
| Ruby district | Magnesian limestone overlain by calcareous limestone | 5,000 feet |
| Preacher district | Volcanic tuff and associated igneous rocks | .................... |
| | Black shale, merging downward into schist | .................... |
| | **Upper Cambrian** | |
| Kandik district | Limestone, with dark gray to black slate and chert in higher part | .................... |
| | Limestone | .................... |
| | **Middle Cambrian** | |
| Kandik district | Upper plate of limestone | .................... |
| | Thin layers of slate and quartzite | .................... |
| | Lower plate of limestone | .................... |

* Stratigraphic succession of the formations listed under any one period is not implied. The relative age is commonly not known.

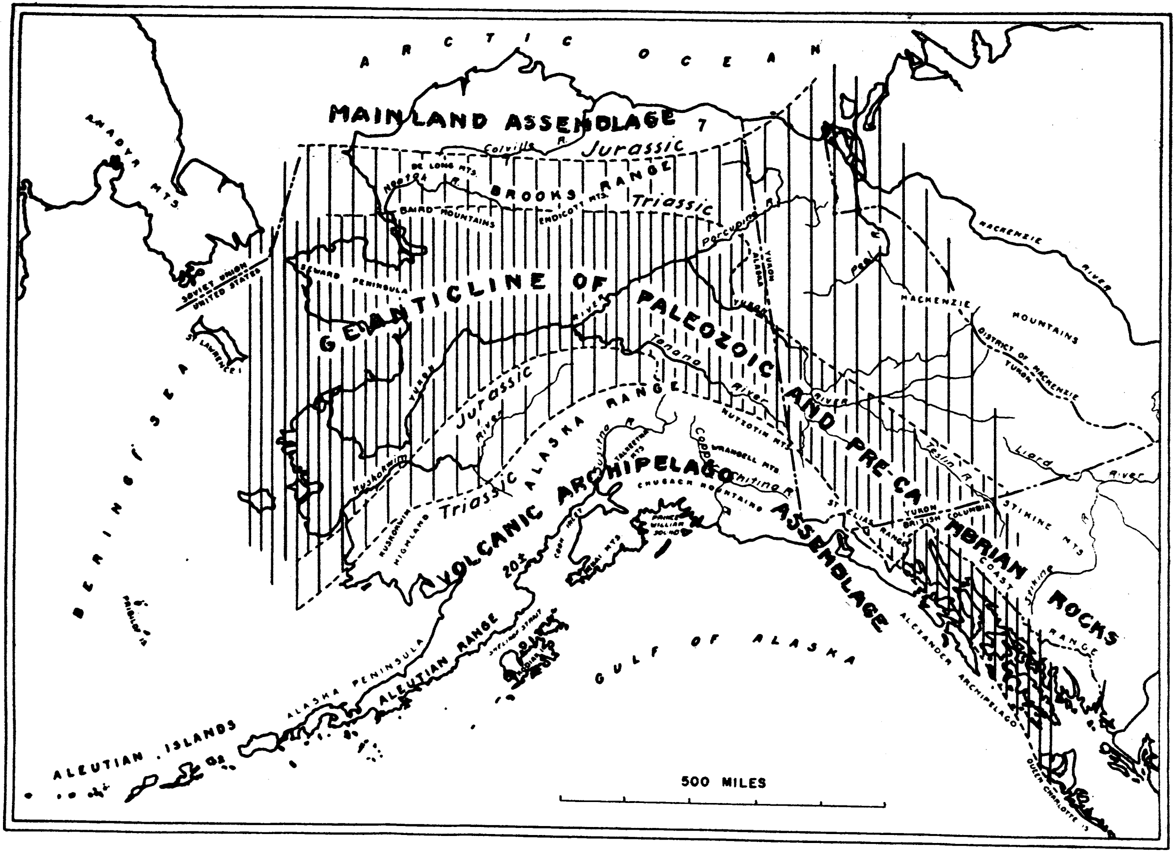



FIG. 3.—Alaskan geanticline of Triassic and Jurassic times and the regions of sedimentation on either side. Vertical ruling covers area of geanticline, which was somewhat farther north in Jurassic time than in Triassic. The close ruling is the area common to both Jurassic and Triassic uplifts, the open ruling on the north shows approximately the limit of the Jurassic uplift, and the open ruling on the south the limit of the Triassic uplift.

cumulation, one on the north and one on the south (fig. 3). The basin on the south collected chiefly sediments of the volcanic assemblage, while the trough on the north received limestones, sandstones, shales, and cherts of the mainland assemblage. Published maps do not exactly delimit the geanticline, but Mertie (1930, pp. 106–110) describes the region of epeirogenic uplift and erosion. The geanticline is analogous, except in detail, to the Mesozoic Cordilleran geanticline in Canada and the United States.

The Mesozoic geanticline that rose in the approximate center of the Paleozoic geosyncline of the Cordillera of the United States and Canada is so characteristic that its presence in Alaska strengthens the concept of an equally broad Paleozoic geosyncline there.

### CRETACEOUS BASINS AND MOUNTAIN BELTS

An examination of the geologic map of Alaska and the correlation chart in Professional Paper 192 shows the Lower and Upper Cretaceous strata so widespread that much of Alaska must have been under water and receiving sediments during Cretaceous times. Certainly, large parts of the Triassic and Jurassic geanticline were covered. Mertie (1930, p. 112) states that at least at one time, perhaps at several, during the Cretaceous, all of Alaska was subjected to sedimentation. But he also concludes, because of the coarse clastic nature of most of the Cretaceous beds and the unconformities at their base and within them, that they were due to "differential warping" and even to mountain building. Linear uplifts and troughlike basins between seem to be discernible if the Cretaceous areas of the Territory are zoned off from the areas devoid of Cretaceous (fig. 4). The Brooks Range probably rose sharply at the end of Lower Cretaceous time and again after Upper Cretaceous time (Mertie, 1930, p. 124). The great Cordilleran geanticline of Canada enters Alaska and appears to split into four prongs, with the Brooks Range constituting the northern one, the Yukon River area from Fort Yukon to Ruby and beyond another, the Kantishna-Kuskokwim Bay belt a third, and the Nutzotin and Alaska ranges the fourth. South of this southern arcuate uplift a long narrow trough formed, in which a series of conglomerates, tuffs, black clays (now slates and phyllites), graywackes, limestones, argillites, and sandstones accumulated. Another uplift bordered this trough on the south which may have connected with the bordering archipelago of southeastern Alaska, British Columbia, and the United States.

Along the north side of the Alaska Range is a group of continental deposits of late Upper Cretaceous age, the Cantwell formation. It has a conspicuous basal conglomerate with boulders up to 6 inches in diameter; and there are many beds of conglomerate throughout the sequence. The formation is several thousand feet thick. It rests unconformably on other Cretaceous beds as well as on pre-Cretaceous rock. According to Mertie (1930, p. 116), the Cantwell was deposited during the first vigorous uplift of the Alaska Range, but the distribution of the pre-Cantwell Cretaceous beds indicates that the range had already been defined by earlier Cretaceous movements.

North of the Brooks Range the Lower Cretaceous has been estimated to be at least 10,000 feet thick, and the Upper Cretaceous at least 15,000 feet (Smith, 1939, p. 55). If they have these thicknesses where superposed, the trough was truly one of geosynclinal proportions. Perhaps enough land lay to the south to

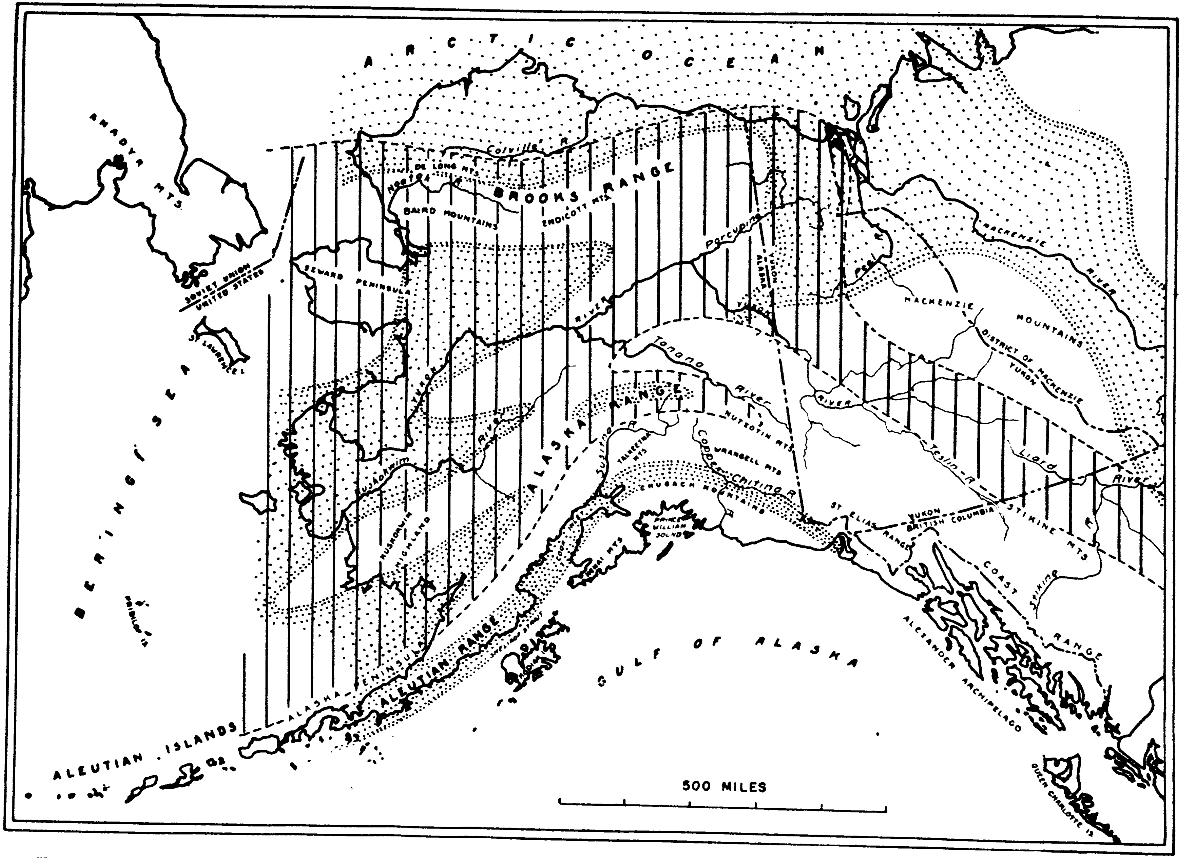



FIG. 4.—Cretaceous seas and linear uplifts and the later Laramide belt of compression. The seas are stippled, the uplifts are blank except where covered by the Laramide belt, and the Laramide belt is ruled. Boundaries are all approximate and based on available information.

furnish all the sediments; but, assuming the distribution shown in figure 4, it seems that at least some doubt is justified. Payne (1948, p. 36) says,

> . . . a thick Lower Cretaceous miogeosynclinal graywacke-shale sequence, without chert and volcanics, formed along the present northern foothills of the Brooks Range. . . . Deposition was rapid and marine in a foredeep between the source in a rising island arc to the south and stable craton to the north.

The Cretaceous in the Lower Yukon embayment is also thick and has been estimated to be over 10,000 feet. Except for the open end toward the Bering Sea, the basin is surrounded by uplifts from which the sediments must have come.

In both the northern Alaskan trough and the Yukon embayment an unconformity between Lower and Upper Cretaceous beds leads Mertie (1930, p. 124) to visualize appreciable crustal disturbance throughout the region at the close of the Lower Cretaceous. According to Payne (1948, p. 37), the Middle Cretaceous orogeny centered in the Baird Mountains of the Brooks Range.

The Cretaceous as a whole is lacking in volcanics except for tuffs; and the volcanoes that emitted the dust for the tuffs probably lay along the southernmost archipelago, which today is the site of the coast ranges or similar elements now under the sea on the continental shelf.

Intrusive igneous rocks are widespread over Alaska, but not enough is yet known to classify them as to age. One group, probably the largest, are Mesozoic in age. In the Alaska Range, in particular, most of them seem to be of late Cretaceous and early Tertiary age (Smith, 1939, p. 90).

### LATE CRETACEOUS AND EARLY(?) TERTIARY OROGENIES

Other orogenic deposits within and north of the Alaska Range, probably younger than the Cantwell and of Eocene age, attest later phases of orogeny. The "coal formation" may overlie the Cantwell unconformably and is itself highly deformed. The Kenai coal-bearing formation south of the Alaska Range contains numerous volcanic rocks, such as breccias, agglomerates, tuffs and flows of basalt, andesite, trachyte, and rhyolite, and suggests crustal unrest; but for the most part no fossils have been found that enable accurate age assignment. It is possible that the volcanic archipelago that confined the Cretaceous trough along the Gulf of Alaska still existed and furnished most of the volcanics to the southern Eocene (?) zone of accumulation.

Above the coal measures north of the Alaska Range is the Nenana gravel, a coarse, well-rounded pebble and boulder deposit, some 2,000 feet thick. Locally it rests unconformably on the coal formation and may be the flanking waste of the newly built mountains. If the gravels are Miocene or Pliocene rather than Eocene in age, then there are two possibilities: first, the orogeny that deformed the coal measures is itself Eocene and was followed by a long period of erosion, after which epeirogenic uplift and the deposition of the gravels took place. On the other hand, the orogeny may be Miocene or Pliocene and may be responsible for deposition of the gravels. These possibilities have been outlined by Mertie (1930, p. 120), who points out also that the Cantwell orogeny was confined to southern Alaska and the Brooks Range but that the second phase (producing the Nenana gravels) spread over the Yukon as well. Both the Upper Cretaceous and the Eocene in the Yukon Basin have been deformed by open folding, and both have been intruded by monzonitic and basic rocks and covered in places by genetically related extrusives.

The site of the Aleutian Range was mostly submerged during the Cretaceous while the Alaskan Range was being outlined, but its late Cretaceous and early Tertiary history is similar to that of the Alaska Range. In later Tertiary times it became involved in Coast Range orogeny.

#### PLIOCENE AND PLEISTOCENE OROGENY

*Areas of deposition and belt of disturbance.*—The southern or Coast Ranges were the sites of the most modern orogeny in Alaska. They consist of Kodiak Island, the Kenai Peninsula, the Chugach Mountains, and the coastal part or more of the St. Elias Mountains. They form a partially submerged, arcuate mountain belt that faces the Gulf of Alaska. The belt consists of Mesozoic strata, largely undifferentiated, and of numerous Laramide (?) intrusions; but numerous Tertiary deposits record its history in late geologic times. Stretching from Lituya Bay to Katalla are outcrops of middle to late Tertiary strata, some sections very thick, and all tilted and folded (Smith, 1939, p. 59).

Quoting from Smith (1939, p. 60):

> As an indication of the former greater extent of these Tertiary rocks the isolated occurrence of them on Middleton Island, far out in the Pacific Ocean, is of particular significance. The rocks on this island are described by Capps as moderately indurated sandstone and conglomerate that in places are inclined 30° or more and have been beveled off by wave erosion, and later gravel and sand have been laid down on the truncated edges.

A thick Tertiary section in the Alaska Peninsula, from Kamishak Bay in the northeast to Pavlof Bay in the southwest, is mostly of volcanic origin. Some of the sediments are probably of Eocene age; others are probably Miocene. They are both terrigenous and marine and in places are probably 5,000 feet thick. Near Herenden Bay, Pliocene beds may occur. In the vicinity of Kodiak Island and on Trinity Island evidence of Miocene and Pliocene beds has been found. All the Tertiary rocks have been folded, dips of the recognized Miocene and Pliocene beds averaging 15°–20° and those of the older Tertiary beds usually much steeper.

By examination of the correlation chart and Geologic Map of Alaska in Professional Paper 192 it will be seen that the younger Tertiary rocks are all confined to the Coast Ranges except some in the Alaska Peninsula. The Coast Ranges were, therefore, the site of subsidence and sedimentation in Miocene and Pliocene time (fig. 5). Beds as young as Pleistocene probably accumulated at the eastern end. Because the strata are tilted and folded, it follows that the same belt became one of mountain building, perhaps during the deposition of the sediments as well as afterward. That the folding was gentler than in the Alaska Range during earlier orogenies is shown by the open character of most of the folds. If the Nenana gravel of the interior of Alaska is of late Tertiary age, it would appear that the arc of the Alaska Range was uplifted at the time of the Coast Range orogeny.

Because the Coast Range orogeny is so recent, the distribution of the mountains indicates the extent of the orogenic belt. The areas of sedimentation of Miocene and Pliocene time, shown by strippling on figure 5, are restricted to the present coastal margins and the continental shelf, whereas the belt of orogeny, shown by ruling, cuts farther inland to include the entire zone of Coast Ranges.

The map of figure 5 shows the relation of the late Tertiary coastal deposits of the Peninsula of Lower California, California, Oregon, Washington, British Co-

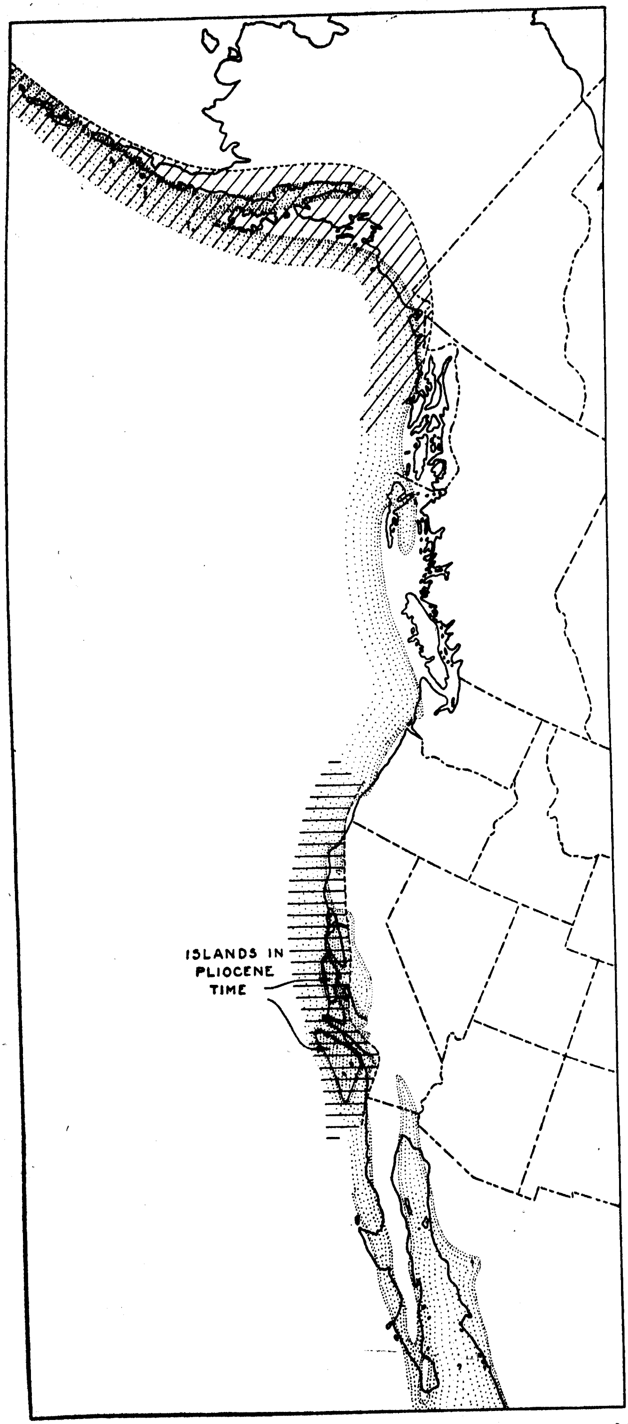



FIG. 5.—Comparison of Alaskan Miocene and Pliocene coastal geology with that of British Columbia, Washington, Oregon, California, and Mexico. Stippled areas represent Miocene and Pliocene deposits, and ruled areas belts of late Pliocene or Pleistocene compressional orogeny, where known. The Alaskan and Californian belts may be continuous but beneath water on the continental shelf.

lumbia, and southeastern Alaska to those of the Coast Ranges of Alaska. Intense compressional deformation in late Cenozoic time was localized in the central and northern Coast Ranges of California (map, fig. 5). The gentle compressive movements which started in the Miocene and then relaxed for a while surged to a peak in the late Pliocene and again to another peak in the mid-Pleistocene. The orogenic belt, if it continued northward, lies seaward of Oregon and Washington and, aside from gentle warping in the Island and Coast Ranges of British Columbia and southeastern Alaska, is not recognized again until the Gulf of Alaska coast is reached at Lituya Bay.

*Modern volcanic belt.*—A great arc of active or only recently dormant volcanoes extends from Mount Edgecumbe on Kruzof Island near Sitka through Mount Wrangell and the whole Alaska Peninsula and the Aleutian Islands to Kiska (fig. 6). This arc is 2,500 miles long, and many of the peaks rise 8,000–11,000 feet above the sea.

Southward from the Mount Spur group at the extreme northeastern limit of southwestern Alaska, the signs of Tertiary to Recent volcanism become increasingly evident until, south of Chignik, they make up practically all the features of the bedrock. The lofty modern volcanoes that overshadow all the other topographic features are dominant in almost every landscape. According to Smith (1939, p. 82):

> Their general features may be summarized by saying that the volcanic activity in the region as a whole seems to have begun somewhat after the beginning of the Tertiary and to have persisted intermittently to the present time. The composition of the lavas has in the main been fairly comparable with that of normal andesites, but more basic phases analogous to basalt and more acidic phases approaching rhyolite are by no means unknown. Practically every kind of volcanic activity, from stupendous explosion to quiet welling forth of lava, is represented in different areas and even in many individual volcanoes.

The arcuate belt of volcanoes is undoubtedly an active orogenic belt and closely related to the late Cenozoic belt of orogeny. Compare the map showing the volcanoes (fig. 6) with the one showing the Miocene-Pliocene deposits and the Coast Range orogenic belt (fig. 5).

### CENOZOIC LAND CONNECTIONS WITH SIBERIA

The Cenozoic history of Alaska is similar in many respects to that of the great western Cordillera of Canada and the United States. It seems certain that, at least during Mesozoic and Cenozoic time, the Cordillera, with its troughs, geanticlines, compressional mountain belts, and volcanic arcs, extended into Siberia and there continued down the coast of Asia with branches inland. Therefore, since the Paleozoic, land and shallow seaways almost continuously connected North America with Asia, and the same conditions probably held during the entire Paleozoic era (Eardley, 1947).

As to the region north of the Brooks Range, now the Coastal Plain and Arctic Sea, it has been referred to by Payne as the "craton," by which he means the central stable region or shield of the continent. It is true that the Brooks Range, the foothill belt, and the Arctic Coastal Plain correspond to our frontal Rockies and Great Plains, but, on the other hand, it has not been proved that any of the sediments of the northern Alaska trough came from the north. They seem to have come from the Triassic and Jurassic geanticline and from the early Brooks Range of Middle Cretaceous age, both on the south. Since marine units are present in all the systems of the Arctic Coastal Plain, it would appear that the

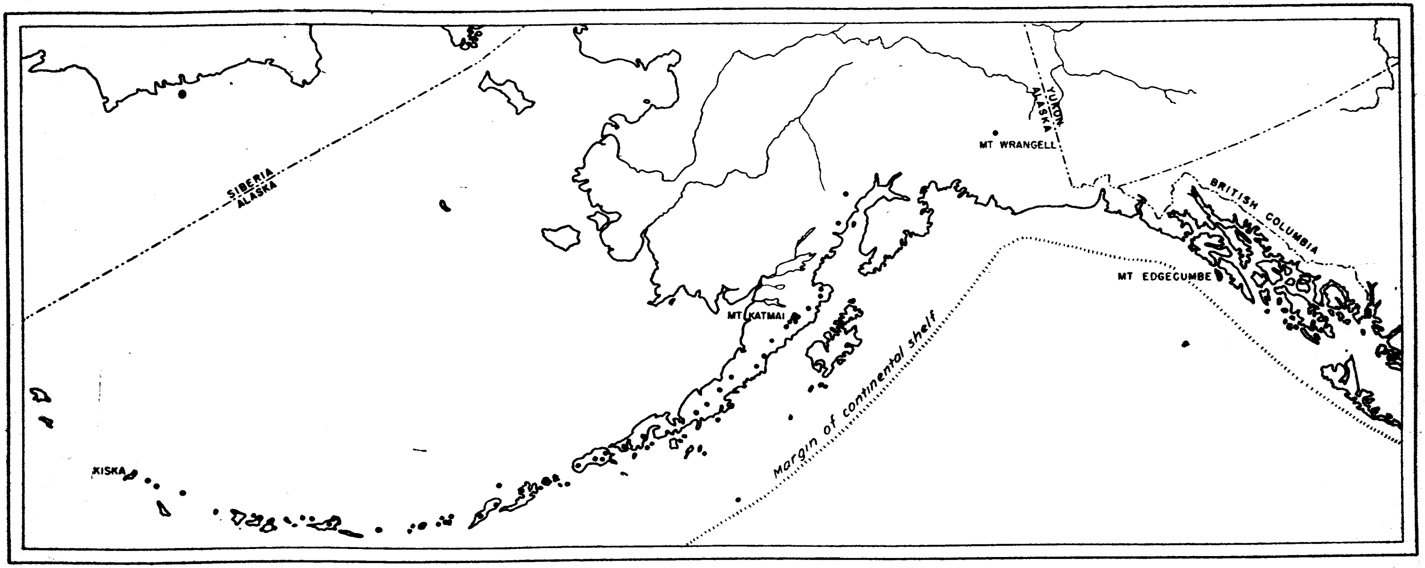



FIG. 6.—Distribution of volcanoes in Alaska. Taken from Smith (1939). Dots are active or recently active volcanoes.

Arctic Sea made its appearance in early Mesozoic time and that its floor continued to subside thereafter. In the next section the geology of the Arctic Archipelago will be considered, and it is there concluded that large parts of the Arctic became seaways in Carboniferous time and that the Arctic Sea basin began to sink then. It does not appear, therefore, that an extensive land can be postulated directly north of Alaska after Carboniferous time. This subject will be considered at greater length later.

## ARCTIC ARCHIPELAGO

### INTRODUCTION AND LITERATURE

The Arctic Islands of Canada, together with Boothia and Melville peninsulas, form a geographic unit known as the "Arctic Archipelago." They extend from Hudson Bay at 62° north latitude northward for 1,500 miles to the northern tip of Ellesmere Island at 83° north latitude. Their greatest extent from east to west is about 1,000 miles.

Little is known of the hinterlands of the Arctic Islands because exploration has been confined, with but few exceptions, mainly to their coasts. However, sufficient information has been gathered to provide a general picture of the geography and geology, though no doubt this will be greatly modified by future more detailed explorations.

The geology of the Arctic Archipelago has been summarized recently by Armstrong (1947, pp. 311–324), and in the following pages the factual information has been taken chiefly from his writing. A report on Victoria Island and vicinity by Washburn (1947) was also helpful. Both reports stress the inadequate information on the vast region. The new Geologic Map of Canada should be consulted for place names used in the following paragraphs.

### PRE-CAMBRIAN AND EARLY PALEOZOIC AREAS

Two large parts of the Canadian Shield, so far as known, are devoid of Paleozoic sediments and made up entirely of pre-Cambrian rocks. They lie on either side of Hudson Bay; the western one is south of Coronation and Queen Maude gulfs and the eastern south of Hudson Strait (fig. 1). Hudson Bay is probably underlain by nearly flat-lying early Paleozoic strata. A vast region to the north in the Arctic Archipelago is made up of both elements, namely, pre-Cambrian rock and flat-lying early Paleozoic beds. If the new Geologic Map of the Dominion of Canada or the Geologic Map of North America is examined, the areas of pre-Cambrian and Paleozoic rocks are seen to be large but irregularly distributed. They include Victoria, Prince of Wales, Somerset, Baffin and Devon islands, and Boothia and Melville peninsulas. In some places the contact of the Paleozoic beds with the pre-Cambrian is exposed, in others the base of the Paleozoic strata is below sea level, and cliffs eroded in the horizontal strata rise precipitously from the shoreline to a height of 1,000 feet. On Baffin Island just north of Cumberland Sound the pre-Cambrian rocks are reported to rise 10,000 feet above sea level. In places Cambrian beds, elsewhere Ordovician and Silurian strata, rest directly on the pre-Cambrian.

It has been pointed out (Armstrong, 1947, p. 320) that, in general, successively younger systems of rocks make their appearance from south to north. On Ellesmere Island there may possibly be a continuous sequence from Cambrian to Triassic or even Jurassic. The Paleozoic rocks also generally increase in thickness from south to north. They range, presumably, from 100 or 200 to 1,000 feet in many coastal areas, but on

Borden Peninsula at the north end of Baffin Island there are about 2,000 feet of Ordovician and Silurian beds. Still farther north in the area of Carboniferous rocks on Ellesmere Island the total Paleozoic section is at least 10,000 feet thick. The Carboniferous rocks on the south side of Melville Island are reported to be 4,000 feet thick. These least two occurrences are mentioned in the next section.

The early Paleozoic strata are dominantly limestones and shales, with sandstone units generally in the Cambrian and Carboniferous systems. This is typically a central stable region or shield assemblage. The overlaps on the pre-Cambrian and the thicknesses also reflect shield geology.

It is tempting to try to define basins and domes or arches in the Paleozoic sediments of the Archipelago. An arch of pre-Cambrian rock may be seen stretching northward along Boothia Peninsula into Peel Sound, and West of the apparent arch is a Paleozoic basin that occupies the southern part of Victoria Island. Elsewhere, however, many scattered remnants of Paleozoic on the pre-Cambrian defy grouping into basins. In none of the areas do the reported thicknesses indicate local basins. If it be concluded that much of the pre-Cambrian was covered with early Paleozoic deposits and that, because of unequal upward movements since, they have been dissected and removed over large areas, it is a mistake to use the present base of the Paleozoic as a guide to Paleozoic basins. In the present state of information, however, it seems best to rest the case on the premise that only a broad division known as "pre-Cambrian and Paleozoic areas" can be recognized. It is surely an integral part of the Canadian Shield.

###### CARBONIFEROUS AND TRIASSIC AREAS

North and West of the area of pre-Cambrian and early Paleozoic rocks is a broad elongate area of Carboniferous strata. They may be followed from Robeson Channel between the northern ends of Greenland and Ellesmere Island southwestward to Dobbin Bay, then westward for about 100 miles, where a large Triassic area is found. To the south on Bjorne Peninsula more Carboniferous rock is reported; then southwestward to Grinnell Peninsula, which is also apparently made up entirely of rocks of the same age. West of this, Bathurst, Melville, Prince Patrick islands, and the northern part of Banks Island are probably underlain mostly, if not entirely, by strata of Carboniferous age.

A large area of Triassic lies west of the Carboniferous on the west coast of Ellesmere Island and the east coast of Axel Heiberg Island. It seems to have a linear nature and is parallel to the general Carboniferous boundary just described. This is especially conspicuous if the small areas of Triassic on Bathurst and Prince Patrick islands are joined with the main mass. The Triassic rocks are sandstones, with "subordinate schists and limestones" (Armstrong, 1947, p. 320). Some of the beds from Scoresby Bay to Cape Cresswell along the east side of the north end of Ellesmere Island may be Triassic, although no fossils have been found in them. They are "slates, quartzites, schists, grits, and limestones" and are known in northwest Greenland as the "Cape Rawson beds."

###### ELLESMERE-GREENLAND FOLDED BELT

Most of the rocks of Paleozoic and Triassic age in the Arctic Archipelago are flat-lying, but in a northeast-southwest belt through Ellesmere Island and the northwest coast of Greenland the beds

are both conspicuously folded and thickened. According to Armstrong (1947, p. 322):

> . . . In the vicinity of Vendome Fiord, on the west coast of Ellesmere Island, folded strata of Silurian and Ordovician age have been observed. A series of northeast-trending, sharp anticlines occurs in this region. Towards the south the flexures become less acute, although in many places, as along the north coast of Baumann Fiord, the strata have a very steep dip. Folding also becomes less severe towards the east and is barely recognizable 11 miles from the head of Makinson Inlet. Little is known of the extent of the folding in other directions. No folds can be seen on the south side of Baumann Fiord where the folded Silurian and Ordovician strata are either buried beneath flat-lying, younger Devonian and Carboniferous strata or swing to the west beneath the sea. If the former is the case the folding is pre-Devonian and probably of late Silurian (Caledonian) age.
>
> Another belt of folded rocks is exposed along the northeast coast of Ellesmere Island north from Scoresby Bay to Cape Cresswell. These folded rocks, called Cape Rawson beds, consist of unfossiliferous slates, quartzites, schists, grits, and limestones. The beds are commonly vertical and have a general southwest trend. Their age is not known. Bentham suggests they are possibly the northeast continuation of the folded Silurian strata on Vendome Fiord, and that the folding at both places is of Caledonian age. Koch, who has studied the Cape Rawson beds in northwest Greenland, also believes the folding is of Caledonian age, though his reasons are not conclusive. Schei suggests that the folding is of post-Triassic and pre-Miocene age, and that some of the Cape Rawson beds are Triassic. He bases this conclusion on the fact that he found lithologically similar folded Triassic rocks on the west side of Ellesmere Island.
>
> Triassic sandstones with subordinate schists and limestones underlie most of the east coast of Axel Heiberg Island and the opposite coast of Ellesmere Island. These rocks have provided diagnostic fossils. Schei states that the dip of the Triassic strata is in many places 50 to 60 degrees, and that folds occur north of Greely Fiord.

The Carboniferous rocks along the south coast of Melville Island near Bridgeport Inlet are known as the "sandstone series" and are at least 3,850 feet thick. They consist of 1,225 feet of shale, with a few sandstone beds, which are overlain by 2,625 feet of white sandstone. The beds have an average dip of 65° south (Armstrong, 1947, p. 319). The coincidence of strike of these beds with the general contact of the Carboniferous rocks and the older strata and the substantial thickness at the locality of high dips suggest that the Ellesmere Island folded belt curves westward through Melville Island. Since very little is known of the geology of Banks Island, the extent west of Melville Island of the folded belt is not known. It is possible that it lies beneath the Beaufort Sea and is expressed by northeast-trending Paleozoic beds in the Tanana-Yukon region of Alaska that Mertie (1935, p. 300) has described.

Because the Triassic beds are folded, it follows that the deformation is either post-Triassic (which would probably be Laramide, judging from Alaska) or in part post-Triassic and in part post-Carboniferous and pre-Triassic. If the latter view is correct, the Triassic beds should be in part coarse, owing to the earlier diastrophism. They are described only as "sandstones" and "schists." The latter word is evidently not used in the American sense, and its meaning is not clear. Possibly the Carboniferous beds overlie the Devonian and older beds unconformably, or possibly an angular unconformity separates the Devonian from the Silurian, and both the Acadian and Caledonian orogenies are represented in the compressional belt. These relations are only conjectural. The coincidence of the belts of Carboniferous rocks and diabase-intruded Triassic with the belt of folding suggests an Appalachian age, at least, for the first main phase of deformation, with

a strong post-Triassic phase definitely indicated.

### TERTIARY DEPOSITS

Outcrops of Tertiary sedimentary rocks with coal beds occur at numerous localities throughout the Archipelago. At the north end of Baffin Island on the south side of Eclipse Sound, on both sides of Navy Board Inlet, and at the head of Isabella Bay interbedded sandstone, shale, and lignite, presumably of Tertiary age, occur. Deposits containing fossil wood crop out in the southwestern part of Prince Patrick Island. Tertiary deposits occur at several places along the northeast and west coasts of Ellesmere Island. They consist of sandstone, shale, and lignite and contain fossil plants. One seam of coal on Stenkuls Fiord is 5 feet thick. Tertiary deposits containing carbonized fossil tree trunks have been found in deposits on the northwest coast of Banks Island.

A Miocene age of the Tertiary deposits is mentioned by Armstrong (1947, pp. 316–324), from whose writing the above résumé is taken. Apparently, Chaney (1940, p. 480) charts them all as Eocene. The Tertiary deposits of the Arctic Archipelago are correlated with the Tertiary of the north coast of Alaska and especially with the "coal formation" of the interior of Alaska, generally thought to be Eocene, although it may be younger. Two coal-bearing series may be present in Alaska, and the same may be true in the Archipelago, but the general Tertiary age of the lignite-bearing strata of the Arctic Archipelago can hardly be doubted. The deposits are widespread and undoubtedly indicate a temperate, moist climate with extensive fresh-water swamps at elevations near sea level. If as young as Miocene, the great climatic change has come to the Arctic in late Cenozoic time. This is discussed under a later heading.

### PLEISTOCENE EPEIROGENY

Washburn (1947) describes well-preserved, raised strand lines and marine fossils, which show that Victoria Island has emerged at least 500 feet. In addition, he believes that the whole of the Arctic Archipelago has undergone comparable uplift. This movement is interpreted as an isostatic adjustment to the unloading of Pleistocene ice. Elevated beaches up to 1,000 feet above sea level occur along the east coast of Hudson Bay (G. M. Stanley, personal communication).

## ARCTIC SEA BASIN

### SHELF AREAS

Soundings in the Arctic Sea areas shown on the Hydrographic Office Chart No. 2560 of the U.S. Navy Department do not include those of the U.S.S.R. North Pole Expedition under the command of Ivan Papanin in 1937 and 1938, and the drift of the "S.S. Sedov" through the polar sea in 1938 and 1939. These Soviet soundings plotted on the U.S. Navy chart, together with others from the North Atlantic and North Pacific, made possible the bathymetric contours of figure 1. A broad shelf area north of Siberia, less than 200 meters deep, is particularly prominent. Much of the sea from Nova Semlya (Novaya Zemlya) and northern Norway to Spitzbergen is less than 200 meters deep and is nowhere as much as 1,000 meters. No deep water has been reported in Hudson Bay or in the Arctic Archipelago except between Baffin Land and Greenland.

### DEEP BASIN

With so much of the area of the Arctic Sea a shallow shelf, the truly deep basin

is not so large as commonly supposed. It is about 1,100 miles wide and 2,200 miles long. The Gulf of Mexico where deeper than 2,000 meters is 300 miles wide and 900 miles long, and the Caribbean Sea is about 400 miles wide and 1,600 miles long. The combined area of the deep basins of the Gulf of Mexico and the Caribbean is somewhat less than that of the deep basin of the Arctic Sea but, as far as size is concerned, the Arctic Sea is much more a mediterranean than an ocean. The Gulf of Mexico and the Mediterranean Sea are considered to be within the framework of the continents and not true oceans, but the Caribbean is thought by some to be bottomed by Pacific Ocean crust (Schuchert, 1935, p. 35).

The few deep soundings that have been made in the Arctic Sea reveal depths of from 3,000 to 6,000(?) meters, with much of the bottom about 4,000 meters deep. Detailed topography is suggested, but the few soundings could be contoured in a number of ways, depending on the interpretation desired.

### PLATFORMS, RIDGES, AND BASINS

The great shelf north of Siberia extends to Alaska and southward beyond the Bering Straits into the Bering Sea (fig. 1). A region connecting Siberia with North America as broad as Alaska is thus under less than 200 meters of water.

The Atlantic Platform, or Mid-Atlantic Rise, extends northward to Iceland, where a transverse, narrow platform stretches from Great Britain to Iceland and to Greenland. Another rather narrow ridge or platform, not anywhere over 1,000 meters deep, connects northern Greenland with Spitzbergen and Spitzbergen with Norway.

There is a great basin on both sides of Greenland. One between Greenland, Spitzbergen, Norway, the Faeroes, and Iceland is irregular but deep, in one place over 4,000 meters. It is about 500 miles wide and 1,300 long. The other basin, that of Baffin Bay on the west side of Greenland, is slightly over 2,000 meters deep over much of its bottom, about 350 miles wide, and about 650 miles long. It is separated by a saddle, a small basin, and another saddle from the main Atlantic, toward which the waters progressively, but irregularly, deepen. The broad saddle area is about 1,000 meters deep.

## GEOLOGY OF LANDS SURROUNDING ARCTIC SEA

### SHIELDS FACING BASINS

*Canadian shield.*—The major tectonic elements around the Arctic Sea are the shields. Aside from the Alaskan orogenic belts and their continuation into the Anadir Peninsula of Siberia, the Arctic Basin is surrounded by the great stable elements of the earth's crust. These shields are separated from each other, or perhaps broken, by intra-shield orogenic belts which project into the Arctic Sea and are covered by it.

The Canadian shield, as explained above, extends to the orogenic belt of Ellesmere Island in the Arctic Archipelago. Beyond the orogenic belt is the undisturbed Carboniferous and Triassic Basin, the beds of which dip gently under the Arctic Sea.

*Greenland shield.*—Around the coast of Greenland are extensive areas of pre-Cambrian rock, and, except for the belt of Caledonian folding along the east side and the late Paleozoic folding along the north, the great island seems to be very similar to the Canadian shield, if not a part of it (Koch, 1929; and Hobbs, 1932, p. 373). Baffin Bay is a fairly deep depression between the two (fig. 1) and, be-

cause of its closed nature, must be of structural origin. This aspect of the geology will be dealt with later.

*Russian-Baltic shield.*—As depicted by Umbgrove (1947, pl. 5), the Russian-Baltic shield extends from the Ural Mountains of the U.S.S.R. westward to the Caledonian orogenic belt of the British Isles, Norway, and Spitzbergen. It, like the Canadian shield, consists of pre-Cambrian rock veneered in part by Paleozoic strata, little deformed except for basins of subsidence. North of the shield is a "massiv" or "nucleus," according to Umbgrove's plates 1 and 2, that has the tectonic character of a small shield. It is called the "Nucleus of Barents Sea and Putkow Kamen."

*Angara shield.*—The great Angara shield of Siberia extends from the Urals to the orogenic belt just east of the Lena River. Probably it was connected with the Russian-Baltic shield until Carboniferous time, when the Ural Mountains were formed in an intra-shield orogenic belt. The eastern edge of the shield is flanked by both Carboniferous and late Mesozoic orogenic belts.

An arcuate Caledonian and Variscan orogenic belt through the Taimir Peninsula and North Land cuts off a piece of the shield and, together with the Variscan belt through Nova Semlya, effectively surrounds the fragment. Umbgrove has called it the "Nucleus of Kara Sea."

*Nucleus of Tschuktschen.*—East of the Lena River is a broad belt of poorly known northward- and northwestward-trending structures which Umbgrove shows to be composed of Variscan and Mesozoic elements. They appear to continue northward through the New Siberian Islands into the Arctic Sea. The eastern part of the New Siberian Islands is pre-Cambrian rock, and so also is Wrangel Island. These exposures, together with the broad shelf, led Umbgrove to depict another nucleus, namely, the "Tschuktschen" flanking a large part of one side of the deep basin of the Arctic Sea.

### PALEOZOIC OROGENIC BELTS

Most of the Paleozoic orogenic belts that occur in the Arctic region have already been mentioned. All are shown on the map of figure 1. The Caledonian belt through Great Britain, Norway, and Spitzbergen is well known. Folds and thrusts along the north and east sides of Greenland are thought to have Caledonian characteristics similar to those of Norway and Great Britain (Koch, 1929, 1935, 1939). The beds are overthrust westward along the east side of Greenland.

The age of the folds of the North Greenland–Ellesmere Island belt is uncertain: they might be Caledonian, Variscan, or even Mesozoic (p. 423).

The Variscan orogenic belt of the Ural Mountains splits before reaching the coast, the right arm extending through Nova Semlya into the Arctic Sea and the left arm northwestward to the coast and under the Barents Sea.

The orogenic belt along the east side of the Angara shield also extends to the coast and runs under the Arctic Sea. Variscan elements are believed to extend along the north shore of the Anadir Peninsula and to curve northward through the New Siberian Islands, where Caledonian folds also have been noted (Umbgrove, 1947, pl. 2). The two extend under the sea. It is possible that the Variscan folds find a continuation in Alaska, but, if so, the evidence is mostly covered and very incomplete.

The possibility of extending the Paleozoic orogenic belts under the Arctic Sea

and connecting them so as to form some kind of logical tectonic pattern is intriguing, but probably several patterns could be devised, and all must be considered a matter of guess. There is no doubt, however, that they project into the Arctic Basin and under the sea. They represent belts of weakness or mobility generally within or between shields and, together with the shields, form a single great tectonic province. It is difficult to escape the conclusion that the shelves, platforms, ridges, basins, and even the deep basin of the Arctic Sea are all part of the province and, more fundamentally, are continental—not oceanic—crustal material.

The great Paleozoic geosyncline and volcanic orogenic belt of western North America and Alaska is of the Pacific border province and distinct from the intra-shield orogenic belts. Alaska is the only approach of the great Pacific belt to the Arctic Sea.

### CARBONIFEROUS BASINS

An attempt is made in figure 7 to show the great lands and the areas of deposition around the Arctic Sea in Carboniferous time. For the American Arctic the new Geologic Map of North America and the new Geologic Map of Canada and the accompanying publication, Economic Geology Series No. 1, were consulted; for Greenland, Koch's (1929, 1935, 1939) various publications; for Spitzbergen, the recent summary article by Allan (1942); for Great Britain, the recent book by Stamp (1947); and for Eurasia, the treatises of Gignoux (1943) and Umbgrove (1947).

A somewhat different map appears when the Carboniferous areas of deposition and the lands from which the sediments came are plotted (fig. 7) from the map showing orogenic belts and shields (fig. 1). In places the earlier Paleozoic orogenic belts had become highlands adjacent to the areas of deposition in Carboniferous time. The Scandinavian-Scottish "continent" seems established (Stamp, 1947, pp. 114–118), but its western extent and relation to the Appalachian geosyncline is unknown. The southern extent of the Carboniferous basin on the east coast of Greenland is a matter of speculation. The southward limit of the Carboniferous basin in the Arctic Archipelago of North America is clear cut, and it need not be projected far eastward to meet the Spitzbergen and eastern Greenland basins or far westward to meet the Cordilleran geosyncline.

Northeastern Siberia is so poorly known that several interpretations of its major tectonic features have been made. The commonest one is that it was all land in Carboniferous time; but this is incongruous with the geology of Alaska, where the great Cordilleran geosyncline existed. From the 1937 geologic map of the U.S.S.R., compiled and printed by the Soviets, it is surmised that the geology is somewhat similar to that of Alaska. This, together with Umbgrove's conception of the "Nucleus of Tschuktschen," has led me to show Carboniferous deposition along northeastern Siberia as well as land to the north. The land seems necessary to supply the mainland assemblage of sediments to the Cordilleran geosyncline in Alaska. This postulated land could have connected with the Canadian shield and not have been cut from it by the Archipelago basin as shown on the map.

The Russian Sea may not have branched into three lanes as shown but may have enveloped the whole broad terrane. At any rate, it appears that the lands and the areas of sedimentation of the northern hemisphere in Carboniferous times were a closely knit province

about the Arctic region and that they stood in contrast to the great Pacific Ocean and its marginal geosyncline and orogenic belts. The Appalachian geosyncline, also a marginal type according to Kay (1947, p. 1290), seems anomalous in the setting if it is projected northerly into the shield province. Its Carboniferous elements have generally been connected with the Armorican belt of orogeny in southern England and Brittany of France, but with great uncertainty.

### MESOZOIC BASINS

Triassic sedimentary rocks north of the folded belt in the Arctic Archipelago have already been mentioned. They dip gently under the waters of the Arctic Sea; their thickness is unknown. When charted on a map of the Arctic region (upper map of fig. 8) they appear to be part of the same basin as the Triassic and Jurassic rocks of the Alaskan Arctic coastal plain north of the Brooks Range. Numerous areas of Triassic and Jurassic rock are shown on the U.S.S.R. Geologic Map of 1937 in the Anadir Peninsula of Siberia across the Bering Straits from Alaska, and presumably they represent the continuation of the same great basin. The Mesozoic rocks continue westward in Siberia to the Lena River and Angara shield and northward through the New Siberian Islands. The distribution shown on the map of figure 8 is based on the work of Arkhanguelsky (1937, p. 299).

Additional Mesozoic sediments occur extensively on the east coast of Greenland (Koch, 1929) and on Spitzbergen (Allan, 1942) (fig. 8). The Jurassic sections in these areas have representatives of most of the epochs. Northern Norway also has Oxfordian and Portlandian strata of the Upper Jurassic. The seas of Callovian time set their marks on Franz Joseph Land and Nova Semlya (Koch, 1929).

Paleogeographic maps of the Arctic can be regarded only as hypothetical. The interpretations of the Siberian Arctic are especially weak. Although our knowledge of Alaska is incomplete, it is sufficient to bring out the incongruous and probably erroneous nature of certain maps that have been published on Siberia across the straits. The only value of an attempt to make a paleogeographic map of the Arctic is to express the belief that the region was no different from other parts of the continents of the Northern Hemisphere. As for the Triassic and Jurassic seas, they transgressed the great shields extensively, and certain lands existed where water now lies. The physical evidence is not enough to postulate land connections between the Eurasian and North American continents in the Greenland-Norway region, but they could have existed at times. The Alaskan connection appears to have been almost continuous because of the geosynclinal and orogenic belt there.

The seas of Cretaceous time were more complex than those of the Jurassic and Triassic, and an attempt to assemble reliable data and to make a map of the seaways led only to the conclusion that it was more a guess than an interpretation, and therefore not worth while. It need only be said that marine Cretaceous strata occur widely and in unexpected places.

### TERTIARY DEPOSITS

Numerous Tertiary deposits have been found in the Arctic region, and fortunately most of them carry coal beds and plant fossils. The lower map of figure 8 is taken partly from a study of the plants by Chaney (1940, pp. 469–488). He attempted to reconstruct the character and

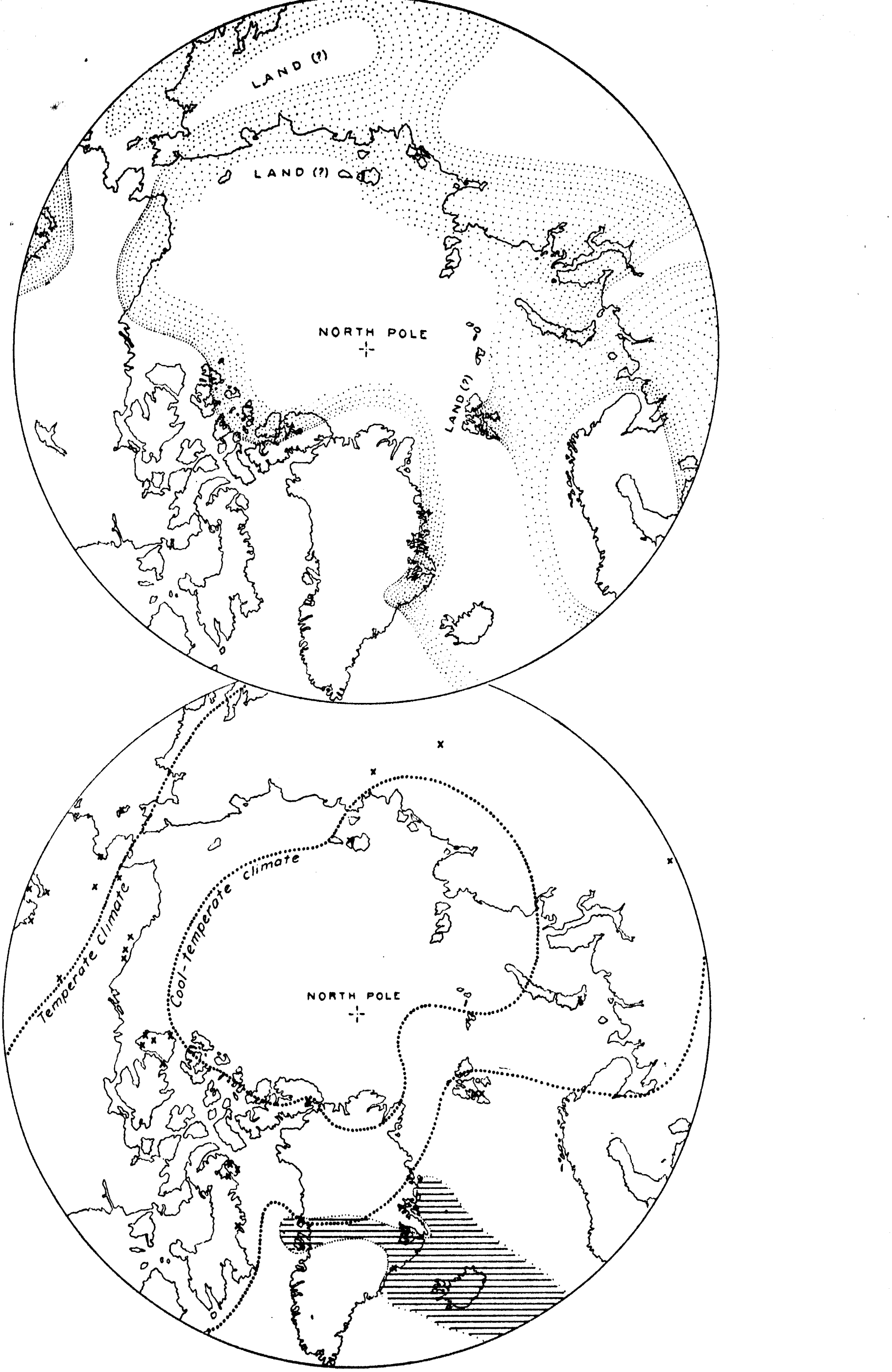



FIG. 8.—*Upper map:* very generalized distribution of seas and lands of the Arctic during Triassic and Jurassic times. The seas at any one time were not so extensive as the total distribution shown.

*Lower map:* early Tertiary deposits of the Arctic. The dotted lines are isoflors after Chaney (1940), and the crosses denote Chaney's Eocene and Oligocene localities, plus a few other localities where "Arctic Miocene" coal beds are known. The ruled area denotes the Greenland-Iceland-Scotland basalt field of early Tertiary time.

distribution of the Tertiary vegetation, in order to learn of the distribution of climate and migration routes in the past. He writes:

> The Tertiary flora which best illustrates a migration response to climatic change has as its dominant element the redwood (*Sequoia*), together with several of its modern associates, alder (*Alnus*), pepperwood (*Umbellularia*), and tan-oak (*Lithocarpus*). An associated element includes members of the modern deciduous hardwood forest of the eastern United States and eastern Asia, basswood (*Tilia*), beech (*Fagus*), chestnut (*Castanea*), elm (*Ulmus*), and hornbeam (*Carpinus*). A third element is made up of general now restricted to Asia, such as katsura (*Cercidiphyllum*), maidenhair tree (*Gingko*), tree of heaven (*Ailanthus*), water chestnut (*Trapa*), and zelkoua (*Zelkova*). Such plants are typically temperate in their modern distribution. Their moisture requirements are high (40 to 60 inches annually), and the predominance of deciduous genera suggests summer as well as winter rainfall. This flora first appeared in the Eocene of high northern latitudes, including Alaska (Kenai flora), (Hollick, 1936) northern Siberia, Spitzbergen, and Greenland. Evidence that some of its genera ranged southward along mountain ranges during this epoch is to be seen in its occasional presence as a minor element in the subtropical floras which characterized middle latitudes in western America and western Europe during the Eocene. The existing forests of Alaska include a number of trees, such as the mountain hemlock (*Tsuga mertensiana*) and aspen (*Populus termuloides*), whose modern distribution southward at successively higher altitudes in the Cordillera corresponds closely to that here suggested for the Eocene redwood forest. Except in the northern portions of their range, where they live at sea level, both of these trees occupy habitats so remote from sites of deposition that their entrance into the contemporary record is rather unlikely. If a future change in climate should lower the temperature and precipitation in the western United States, the forests now living in our lowlands would give way to vegetation from the north and from the Cordillera. This northern and alpine vegetation would leave a record as soon as its altered range brought it down into the valleys, in close proximity to sites of deposition.

Chaney's conclusions are that the climatic zones were distributed around the North Pole in early Tertiary time with modifications due to land and warm ocean currents, as today, except that the rain-forest climate (temperate) of the California Coast Ranges was near the pole, and subtropical conditions existed in California, or as far north as latitude 40; that the factors controlling air and water circulation were essentially the same then as now. He believes that the gradual cooling of the Arctic region is coincident with a gradual uplift of the continent and that there was a land connection between Siberia and Alaska almost continuously during the Tertiary.

Figure 8 also shows the extent of the great basalt field between Greenland and Scotland. It is possible that much land existed in this area during the eruptions and afterward.

### TERTIARY CRUSTAL MOVEMENTS

Following the deposition of the Tertiary coal-bearing beds, the region of Spitzbergen was gently uplifted in the eastern and central districts, but steep folding and overthrusting occurred along the west. The large island was taking its modern shape, and, according to Allan (1942, p. 46),

> Stor Fjord was a major depressed fjord block and Ice Fjord with its branches was probably due to a combination of east-west tear-faulting and associated block faulting. Today the etched and buttressed slopes of the horizontally bedded rock masses flanking Ice Fjord display numerous examples of step faulting toward the shore lines. The Tertiaries on either side of the Foreland Sound dip toward it—yet another trough faulted feature.

Rather gentle folding occurred in Britain in the Miocene and reflected the superior mountain building in the Alpine chains of the continent, and by the close of the Miocene the structure was vir-

tually complete. At the beginning of Pliocene time southeastern England was 650 and 700 feet lower than today (Stamp, 1947), and, while the land rose, a succession of platforms was carved by the waves. One of the best is the "400 foot platform"—400 feet above sea level.

Faults have been recognized on both the west and the east coasts of Greenland. The Cape York district of northwest Greenland is especially broken by high-angle faults (Koch, 1929), and the fiords of the west coast about Disko and Umnak bays generally have their courses parallel to faults (Hobbs, 1932, p. 380). It is not clear, however, that these faults were associated with Tertiary land movements. Koch (1935) believes that strong Tertiary faulting may be recognized in many places along the eastern coast and that it was associated with the great volcanic activity. The faults have tilted a plane to the west on Milne Land and may be seen cutting the sediments there. Along the east side of Hurry Inlet are Tertiary faults, and Liverpool Land was doubtless strongly raised in Tertiary time. On Wegner peninsula and in nearby fiords, on the east side of Traill Island, on Geographical Society Island, on Ymer Island, and several others north to the Wollaston Foreland, many faults have been recognized and are believed to be of Tertiary age. Part of the movements predate the great lava flows, and part postdate them. The volcanics of eastern Greenland are several thousand meters thick in places and, as a number of writers have proposed, must be continuous with the basalt fields of Iceland, the Faeroes, and Scotland.

The geology of northern Siberia is still very obscure, but a large marine transgression from the north occurred in Quaternary time, and much of the geology of this vast region is hidden under the blanket of the sediments of this sea and also certain glacial deposits. The Khatanga depression along the Khatanga River, which cuts the northern part of the Angara shield, has been considered to be bounded by faults and to be of Tertiary age. According to Shanazarov (1948, p. 172):

> . . . Epeirogenic phenomena took place during the entire Tertiary and Quaternary, bringing about the sinking of some blocks of Siberian territory and the raising of others. According to recent observations these epeirogenic movements are continuing even to the present.

Although the Arctic localities where Tertiary faulting and land movements have been observed are relatively few and the details and ages are not precisely known, they do suggest crustal movements that could have affected profoundly the distribution of land and sea. It seems entirely possible that faulting and epeirogenic movements somewhat comparable to those of the Great Basin in western North America in mid- and late Tertiary time could have occurred and that the basins of Baffin Bay, the Greenland Sea, and even in part the deep Arctic Sea basin could have been formed in somewhat their present shape during the Tertiary. The great thickness of basic eruptives in the Scotland-Greenland region must have altered the land areas considerably there also.

### PLEISTOCENE EPEIROGENY

Isostatic uplift due to the melting of the glaciers, of 600 feet in Victoria Island and of 1,000 feet in Hudson Bay, has already been mentioned. The Scandinavian Peninsula was depressed by the last ice and has risen 600 feet since its dissipation. In all places the rise is believed to be continuing.

If the ice now on Greenland were to melt, the land there would rise about 600

feet if the Scandinavian region is taken as a precedent. According to James W. Wilson's isostatic computations for the writer, the center of Greenland would rise 600–1,000 feet and the uplift would extend beyond the present shores but not enough to connect Greenland with Spitzbergen or Iceland. It probably would merge Ellesmere Island with Greenland (fig. 1).

If the fossil record should demand major land connections between Greenland and Europe during the Pleistocene, it is evident that movements other than those caused by loading and unloading by the ice must be sought. These seem to be a continuation of the late (?) Tertiary movements just described that undoubtedly are of deep-seated origin.

## SUMMARY OF EVIDENCE OF ANCIENT ARCTICA

A review of Alaskan sediments has led to the belief that the Paleozoic Cordilleran geosyncline of western North America extended through Alaska and embraced the whole of the territory. The Pacific side of the geosyncline is marked by the volcanic-bearing assemblage. Such an arrangement requires an adjacent land on the north, where sea now exists. This observation was the first that directed attention to the theory that the Arctic Sea was not an ocean with a permanent basin but that the region had been a land area at times in the past and that Eurasia and North America were broadly connected.

The next observation that supports this concept of Ancient Arctica is the study of the topography of the surface below water in the Arctic region. Much of the sea floor is very shallow and the truly deep basin is only slightly larger than the combined deep-water basins of the Gulf of Mexico and the Caribbean. The parts that range in depth from the shallow shelves to the deep Arctic Sea floor are marked by basins, escarpments, and ridges; and it seems probable that, when bottom profiles are made with the sonic depth finder, that much more detail will come to light. The soundings made of the deep Arctic Basin are pitifully few, and it does not seem logical to conclude that the floor is as smooth as the contours of figure 1 indicate. The depths in excess of 12,000 feet might be due to faulting; and the depression, when better outlined, might have graben characteristics. The presence of considerable topographic detail in much of the surface now covered with water in the Arctic region suggests that the crust there is composed of continental material, and, therefore, it is believed that the North Polar waters should be classed as a mediterranean sea—not an ocean.

The third supporting observation is the fact that the shields of the Northern Hemisphere all face the Arctic Sea and, except for Alaska, surround it with a common geology. It looks as if the deep basin were only a sunken part of a single, great continent.

The shields are separated by several Paleozoic orogenic belts that project to the Arctic Sea and, although in full development at its margin, are lost under it. They surely project long distances into the region now covered by water.

A study of the areas of sedimentation in Paleozoic, Mesozoic, and Cenozoic time on the lands around the Arctic Sea shows clearly that the epeiric sea conditions, which prevailed there at a number of times, were very similar to those of the central stable region and the Canadian shield of North America. Certain of the great seas that invaded North America started in the Arctic and progressively spread southward. Numerous Tertiary

deposits containing coal are known in the Arctic lands. They attest temperate, moist conditions in the Arctic, very unlike those of today, even as late as the Oligocene. Land was evidently more extensive at the time that the coal swamps existed than now, because the beds are now mostly submerged.

A sixth important observation that bears on the continental makeup of the Arctic is the evidence of faulting and major epeirogenic movements in Tertiary time. The late (?) Tertiary thrust faulting on Spitzbergen, the numerous high-angle faults in the marginal areas of Greenland, Pliocene uplift in Britain, and the faulting in the Khatanga depression and elsewhere in northern Siberia all betray widespread crustal unrest in Cenozoic time throughout the Arctic. It is not difficult to visualize these movements as having been of the magnitude of the late and mid-Tertiary faulting in the Basin and Range Province of western North America. As Britain had acquired its present structure and form by the close of Miocene time and as Spitzbergen took its present outline during the post-Miocene (?) faulting, so also it appears that the topography of the Arctic Sea floor, together with that of Baffin Bay, the Greenland Sea, and the North Atlantic, acquired its modern form at about the same time.

A major lava field or volcanic province stretches from Scotland to Greenland but is now covered largely by water. The volcanic rocks, early Tertiary in age, are very thick in places and perhaps at times effectively bridged Europe with America.

As an eighth observation, the isostatic uplift in response to the melting of the continental glaciers has not yet run its course, and a little more land will yet appear because of the continued rise. If the ice of Greenland were to melt, a rise of about 600–1,000 feet would occur, which would drain the water between Greenland and Ellesmere Island but not connect Greenland with Spitzbergen or Iceland.

As a ninth and last observation, the frequent, widespread distribution of animals and plants common to both Eurasia and North America requires migration routes that were probably continuous across the Pacific and common across the Atlantic. It appears to the writer that the geology of the Arctic permits such migration and that the implied connections are much more natural and plausible than a concept of long, narrow land bridges across the Atlantic or the drift of the continents.

### PALEONTOLOGIC SIGNIFICANCE OF ANCIENT ARCTICA

Much has been written about the similarities of Appalachian trilobites with those of Britain and the common varieties of Spitzbergen and North America brachiopods. The realization was startling long ago that Asiatic cephalopods had counterparts in western North America. These seem simply explained by the migration routes that the shifting epeiric seas and orogenic belts of Paleozoic and Mesozoic Arctica afforded. Because of the warm water that the fossils of the far north indicate, climatic barriers to migration could not have prevailed.

Simpson (1947, pp. 613–688) has recently reviewed the genera and families of mammals common to Eurasia and North America and has ably examined the significance of faunal resemblance. He concludes:

> Major faunal interchanges occurred in early Eocene, late Eocene, early Oligocene, late Miocene, middle and late Pliocene, and Pleistocene. There was little or no interchange in the

> middle Eocene and middle and late Oligocene. Each interchange involved migration in both directions, but there was probably more movement from Eurasia to North America than in the opposite direction. All interchanges were selective, and they became increasingly limited from Eocene to Pleistocene. The most important selective influence was probably the relatively cold climate of the land connection. This connection was probably from Siberia to Alaska throughout the Cenezoic and was in almost continuous existence with important interruptions in parts of the Eocene and Oligocene and perhaps shorter interruptions at later times.

Simpson also points out that the Bering connection is adequate to explain all known distributions and that an Atlantic connection is not necessary but still not ruled out. He also cites an oral communication from Colbert to the effect that the migration route of Cretaceous dinosaurs from central Asia to North America was Pacific, not Atlantic.

The bearing of the fossil plants of the Tertiary on climate and land distribution has already been discussed.

It is evident that the mammalian fossil evidence points out more times of land connections than the physical geology yet reveals. The physical geologist has only the Tertiary coal-bearing beds of the northern regions and their deformation as time markers to guide him. The Tertiary beds have generally been called "Arctic" Miocene, but most of them are probably late Eocene, and the subsequent orogeny, such as on Spitzbergen, may be Alpine. When more complete study has been made of the Tertiary beds cropping out in the Arctic, they may prove to be of several ages, and then more can be said about the physical history. Conclusions that may be drawn in consideration of both the fossil and physical evidence are as follows:

1. The Alaskan-Siberian region is one of extensive, very shallow water today and has been the site of geosynclinal and orogenic activity of the Pacific marginal type from early Paleozoic times to the present and has been, therefore, a region of frequent island arcs, mountain chains, and broad arches. As such it would have served effectively and persistently as a migration route of both land and shallow-water faunas. Here the organic record corresponds well with the physical.

2. Migration routes in pre-Carboniferous time, especially, need not have been limited to the geosynclinal belt but could have followed the lands and the stable region of what is now the Arctic Sea and the shores of the epeiric seas that invaded it from time to time. The scant data now available suggest that the deep basin of the Arctic Sea began to form in Carboniferous time and that the principal land migration routes thereafter circumvented it.

3. Although, according to Simpson, an Atlantic "bridge" is not necessary, the physical evidence indicates that connections could have existed. Judging from the topography of the sea floor, two restricted routes were the most recent, one from Asia and Scandinavia via Spitzbergen and northern Greenland and one from Great Britain via Iceland and central Greenland. It is also not inconceivable that extensive land areas connecting Greenland with Eurasia existed.

## PETROLEUM POSSIBILITIES

### PREVIOUS RECOGNITION

The Arctic region has not been overlooked in the search for petroleum. In a recent article Pratt (1947, pp. 658–659) accords it a promising future on three counts: first, it is an area of widespread continental shelves; second, the Arctic Sea is the world's "fourth mediterranean" and is still unexplored, whereas some

of the world's richest reserves are clustered about the other three; and, third, surface evidences of petroleum are known there.

The extent of shallow sea in the Arctic can be determined approximately by inspection of the map of figure 1, and it is true that the amount is great. The concept that the Arctic Sea is a mediterranean has already been elaborated, and it is pleasing to note that Pratt arrives at the same conclusion. It is largely an intra-shield mediterranean, however, and not everywhere associated with Cenozoic orogenic belts, like the others.

The Arctic coastal plain and foothills to the Brooks Range are already being explored, and successful results are expected (Reed, 1946, pp. 1441–1443).

Greenland presents only remote oil possibilities because of its pre-Cambrian terrane, its Caledonian orogenic belts, its Tertiary lavas, and, finally, its great icecap.

Spitzbergen, Nova Semlya, North Land, and the New Siberian Islands are in large part Paleozoic orogenic belts and would normally not interest the petroleum geologist, although bituminous limestones and shales are reported on the New Siberian Islands (Shanazarov, 1948, p. 183). Fohs (1948, pp. 317–330) and Shanazarov (1948, pp. 153–197) have examined literature on northern Siberia for petroleum possibilities. Of the Arctic areas in Siberia the Khatanga depression seems the most promising, although largely unknown and difficult to explore. The mountains of the vast area of northeastern Siberia and the Anadir Peninsula are generally unknown geologically, but vast areas are underlain by marine Mesozoic deposits which at places seem to offer favorable structural conditions for the accumulation of petroleum (Shanazarov, 1948, p. 183).

#### ARCTIC ARCHIPELAGO

Very little attention has been accorded the Arctic Archipelago for its oil possibilities, although it seems to the writer that it is one of the most promising regions of the Arctic. North of the Greenland-Ellesmere folded belt there appears to be a foothill and coastal plain province of Carboniferous, Triassic, and, perhaps, Jurassic rocks, which extends through the northern part of Ellesmere Island, the Sverdrup Islands, Cape Grinnell Peninsula, Barthurst Island, Melville Island, Prince Patrick Island, Borden Island, and the northern part of Banks Island (examine maps, figs. 1 and 8, and the Geologic Map of North America, 1947). Although little explored, the strata are the usual epeiric sea sediments with some intercalated fresh-water coal beds. They are flat-lying or gently folded north of the folded belt. Actual oil seeps of petroleum or bitumen have been reported on northern Melville Island (Armstrong, 1947, p. 320). Such a region possesses the geologic requisites of an oil province. It is not known whether the seeps are from Triassic or Carboniferous strata, but, if from Triassic, the Carboniferous beds will be found at depth. Altogether, there are about 100,000 square miles of land in that part of the Archipelago underlain by Carboniferous strata, and they are worth careful investigation. The deterrent to detailed study is the Arctic climate; but, with airborne support, geologists could do a great deal of profitable field work in the summer months. In addition to hindering field work, the Arctic climate would make the delivery of oil by ship very difficult; but if the need were sufficiently great it seems possible that tankers could get in and out during August and September, at least, of each year. A study of the elaborate publication, "Ice Atlas of

the Northern Hemisphere" by the Hydrographic Office of the U.S. Navy Department, indicates that the unnavigable polar pack impinges on Prince Patrick Island, Borden Island, and the northern coast of Ellesmere Island the year round; but, in the months of August and September, Davis Strait, Baffin Bay, and Lancaster Sound are open, and a passage westward through Melville Sound to Beaufort Sea generally can be made by heavily built vessels without the aid of an icebreaker. It seems probable that contact with shore storage facilities could be made on any of the islands during at least two months of the summer, except along the shores facing the polar pack, if heavy icebreakers were used.

ACKNOWLEDGMENT.—The writer is especially appreciative of the ready help of Professor W. H. Hobbs, whose knowledge of the Arctic literature and geology and whose personal library have made the task of preparing this article a very pleasant one.

## REFERENCES CITED


ALLAN, D. A. (1942) Geology of Spitzbergen: Liverpool Geol. Soc. Proc., vol. 18, pts. 2 and 3, pp. 37–48.

ARKHANGUELSKY, A. D. (1937) Structure géologique et histoire géologique de l'URSS: 17th Internat. Geol. Cong., vol. 2, pp. 285–304.

ARMSTRONG, J. E. (1947) The Arctic Archipelago. In Geology and economic minerals of Canada: Econ. Geology Ser. 1, pp. 311–324.

CHANEY, R. W. (1940) Tertiary forests and continental history: Geol. Soc. America Bull. 51 pp. 469–488.

EARDLEY, A. J. (1947) Paleozoic Cordilleran geosyncline and related orogeny: Jour. Geology, vol. 55, pp. 309–342.

FOHS, F. J. (1948) Petroliferous provinces of Union of Soviet Socialist Republics: Am. Assoc. Petroleum Geologists Bull. 32, pp. 317–350.

GIGNOUX, MAURICE (1943) Géologie stratigraphique, Mason & Cie, 3d ed.

HOBBS, W. H. (1932) Greenland, the advances of a decade (1921–1931): Michigan Acad. Sci., Arts and Letters Papers, vol. 18, pp. 363–411.

KAY, MARSHALL (1947) Geosynclinal nomenclature and the craton: Am. Assoc. Petroleum Geologists Bull. 31, pp. 1289–1293.

KOCH, LAUGE (1929), Stratigraphy of Greenland: Meddelelser om Grönland, vol. 73, pp. 205–320.

——— (1935) Geologie von Grönland, Geologie der Erde, Gebrüder Borntraeger.

———, *et al.* (1940) Grönland, 1939: Naturf. Gesell. Mitt., vol. 16.

MERTIE, J. B., JR. (1930) Mountain building in Alaska: Am. Jour. Sci., vol. 20, pp. 101–124.

——— (1935) Pre-Cambrian and Paleozoic volcanism of interior Alaska: Am. Geophys. Union Trans., 16th Ann. Meeting, pt. 1, pp. 292–302.

PAYNE, T. G. (1948) Tectonic and sedimentary setting of Naval Petroleum Reserve No. 4, northern Alaska (abstr.): Abstracts of Papers Presented at the Annual Meeting of the Am. Assoc. Petroleum Geologists, Denver, pp. 36–37.

PRATT, W. E. (1947) Petroleum on continental shelves: Am. Assoc. Petroleum Geologists Bull. 31, pp. 657–672.

REED, J. C. (1946) Recent investigations by United States Geological Survey of petroleum possibilities in Alaska: Am. Assoc. Petroleum Geologists Bull. 30, pp. 1433–1443.

SCHUCHERT, CHARLES (1935) Historical geology of Antillean-Caribbean region, New York, John Wiley & Sons, Inc.

SHANAZAROV, D. A. (1948) Petroleum problem of Siberia: Am. Assoc. Petroleum Geologists Bull. 32, pp. 153–197.

SIMPSON, G. W. (1947) Holarctic mammalian faunas and continental relationships during the Cenozoic: Geol. Soc. America Bull. 58, pp. 613–688.

SMITH, P. S. (1939) Areal geology of Alaska: U.S. Geol. Survey Prof. Paper 192.

STAMP, L. D. (1947) Britain's structure and scenery, London, Collins.

UMBGROVE, J. H. F. (1947) The pulse of the earth, The Hague, Martinus Nijhoff.

WASHBURN, A. L. (1947) Reconnaissance geology of portions of Victoria Island and adjacent regions, Arctic Canada: Geol. Soc. America Mem. 22.

# SOME MELILITE SOLID SOLUTIONS[1]

JULIAN R. GOLDSMITH
University of Chicago

ABSTRACT

The compositional problems, especially the soda content, of the melilite group of minerals are reviewed. The system $Ca_2Al_2SiO_7$ (gehlenite)–$Na_2Si_3O_7$ was investigated; 15 per cent of $Na_2Si_3O_7$ was found to go into solid solution with gehlenite. Akermanite forms no solid solution with $Na_2Si_3O_7$. A portion of the system gehlenite–$CaO \cdot 2Al_2O_3$ was investigated to determine whether any solid solution takes place. Evidence, not decisive, suggests that gehlenite may take approximately 10 per cent of the calcium aluminate in solid solution.

## INTRODUCTION

The melilite group of minerals presents compositional problems that several workers have attempted to resolve. The complete solid solution series between gehlenite ($Ca_2Al_2SiO_7$) and akermanite ($Ca_2MgSi_2O_7$) is now well known, and it is generally agreed that natural melilites are composed principally of mix-crystals of these two components. The iron analogue of akermanite, $Ca_2FeSi_2O_7$, has been synthesized (Bowen, Schairer, and Posnjak, 1933), and a zinc melilite, Hardystonite ($Ca_2ZnSi_2O_7$), is known. The iron and zinc found in melilites are thus easily explained. Many chemical analyses, however, show rather large amounts (up to a maximum of 6.11 per cent) of $Na_2O$ and $K_2O$, and silica in excess of the values accounted for by mixtures of the akermanite-gehlenite "molecules."

A. N. Winchell (1924) took up the melilite problem from the viewpoint of volume isomorphism and largely discredited the sarcolite molecules which had been proposed earlier by W. T. Schaller (1916) and partially verified by A. F. Buddington (1922). Winchell pointed out the essential $R_5O_7$ character of the melilites and suggested $Na_2Si_3O_7$ and $Ca_3Si_2O_7$ as the most likely "molecules" entering the melilite structure. The soda and excess silica observed in some melilites would thus be accounted for, and excess lime (which Winchell thought existed in some melilites) would also be explained. In addition to the above hypothetical end-members, Winchell also suggested another means of explaining excess silica by supposing that $SiO_2$ is "interatomically" dispersed in the structure. This view was based on the concept, now known to be erroneous, that the oxygen atom was very small. Three more complex molecules were also considered as possible end-members; but there are structural objections to all of them, as in the case of interatomic $SiO_2$, and they will not be considered here.

Berman reviewed the problem in 1929 and modified Winchell's fundamental $R_5O_7$ to $X_2Y_3O_7$. He discarded all the end-members suggested by Winchell with the exception of $Na_2Si_3O_7$. Berman objected to the $Ca_3Si_2O_7$ "molecule" because of its 3:2:7 ratio (not found in natural melilites) and because, if this molecule is used as a basis for recalculation of melilites from the twenty-three chemical analyses he collected, there appears an apparent excess of as much as 9 per cent of silica. Recomputation of melilite compositions from the available analyses, using various end-members, led Berman to replace Winchell's calcium

[1] Manuscript received February 24, 1948.

silicate with his "submelilite" molecule, $CaSi_3O_7$. The deficiency of $CaSi_3O_7$ in lime was used to account for the observation that many of the analyses do not show the theoretical Ca + Na total of 20, if oxygen is set at 70 (i.e., there is a deficiency in the $X_2$ value). Berman also interpreted the $CaSi_3O_7$ molecule as indicating that part of the melilite structure is open. The end-members postulated by Berman are:

Gehlenite —$Ca_2Al_2SiO_7$
Akermanite —$Ca_2MgSi_2O_7$
Soda melilite—$Na_2Si_3O_7$
Submelilite —$CaSi_3O_7$

Recalculation of the chemical data led Berman to state that in the natural occurrences the submelilite molecule does not exceed 10 per cent and the soda melilite does not exceed 25 per cent.

It is unlikely, however, that the submelilite "molecule" can go into solid solution in the melilites; and the same may be said for the $Ca_3Si_2O_7$ "molecule" of Winchell. Aside from difficulties inherent in the replacement of Ca for Al in gehlenite that would be required in the latter formulation, the work of Schairer and Osborn (1941) on the systems Gehlenite–$CaSiO_3$ and akermanite–$CaSiO_3$ disproves the existence of either of the two end-members in the melilites. Both systems are truly binary and show no solid solution. If *any* calcium silicate entered either akermanite or gehlenite, the above systems would show solid solution and would not be binary, as the Ca and Si would be removed from the melt in the proportions necessary to form mix-crystals, resulting in a nonbinary residuum. In addition to this rather conclusive evidence, it is doubtful that the melilite structure has the "holes" that would be required by Berman's formulation of $CaSi_3O_7$.

In 1930 Warren and Machatschki published papers on the melilites. Warren first pointed out, on the basis of X-ray evidence, that the general formula should be $(Ca, Na)_2(Mg, Al)_1(Si, Al)_2O_7$, as there is no evidence to indicate that Si and Mg can ever substitute for each other. Machatschki discussed the same point on the basis of the considerable size difference between Si and Mg and wrote the melilite formula as $X_2YZ_2(O, OH)_7$, where

$$X = Ca, Na, (K)\,,$$
$$Y = Mg, Fe^{++}, Fe^{+++}, Al\,,$$
$$Z = Si, Al\,.$$

This formulation, which doubtless is essentially correct, is the first in which the possibility of OH groups entering the structure is recognized. Machatschki also pointed out difficulties inherent in chemical analyses based on small samples and suggested that the small size and impurity of the samples might explain the low Ca + Na values often found. An alternative suggestion was that the larger cations (K, Na, Ca) might be leached out by secondary chemical action without greatly disturbing the structure; this, however, is unlikely in the case of the melilites, since removal of these ions would result in structural disruption.

## THE SODA CONTENT OF MELILITES

### THE SYSTEM $Ca_2Al_2SiO_7$ (GEHLENITE)–$Na_2Si_3O_7$

$Na_2Si_3O_7$, named "soda melilite" by Berman, does not exist as an independent compound. This fact, of course, does not preclude soda and silica from entering into solid solution as a component in the 1:3 ratio expressed by $Na_2Si_3O_7$. To determine the extent of soda substitution in gehlenite, the system $Ca_2Al_2SiO_7$–$Na_2Si_3O_7$ was investigated.

The technique of investigation, devel-

oped at the Geophysical Laboratory, is now so well known that it will not be here described. Homogeneous glasses prepared from mixtures of pure $SiO_2$, $Al_2O_3$, $Na_2CO_3$, and $CaCO_3$ were subjected to thermal study by the quenching method. Figure 1 represents the results of this investigation. The thermal data from which the equilibrium diagram was constructed are listed in table 1.

$Na_2Si_3O_7$ goes into solid solution with gehlenite to the extent of 15 per cent by weight, equivalent to 3.85 per cent of $Na_2O$. The extent of the solid solution was determined by observation of the completely devitrified mixtures; nonhomogeneous crystallizations result when the amount of $Na_2Si_3O_7$ exceeds 15 per cent in the mix. The limit of solid solution can also be clearly determined with the aid of figure 2, which is a plot of index of refraction (Na vapor light) against percentage of $Na_2Si_3O_7$. Included on this plot are the indices of the glasses in the system. The indices of pure gehlenite ($\omega$ = 1.669; $\epsilon$ = 1.658) drop to values of $\omega$ = 1.644; $\epsilon$ = 1.628 at 15 per cent $Na_2Si_3O_7$, at which point the limiting index values expressed by a break in the curve indicate the limit of solid solution.

Inasmuch as the composition $Na_2Si_3O_7$ does not crystallize as an independent compound but forms $SiO_2$ and $Na_2Si_2O_5$ (Kracek, 1939), the system ceases to be binary at solidus temperatures beyond the solid-solution limit. The diagram was for this reason not extended beyond this point.

### THE SODA-AKERMANITE AND GENERAL SODA-MELILITE RELATIONS

To obtain additional evidence on the presence of soda in the melilites, the relation between akermanite and $Na_2Si_3O_7$ was experimentally investigated. It was found that akermanite takes no $Na_2Si_3O_7$ in solid solution. This statement is based on the results obtained from one mix, of the composition 10 per cent $Na_2Si_3O_7$–90 per cent $Ca_2MgSi_2O_7$. This mix, if devitrified completely (i.e., below the solidus), consists of crystals of pure akermanite and a fine-grained residual crystalline aggregate, undoubtedly composed of $SiO_2$ and $Na_2Si_2O_5$, although no attempt

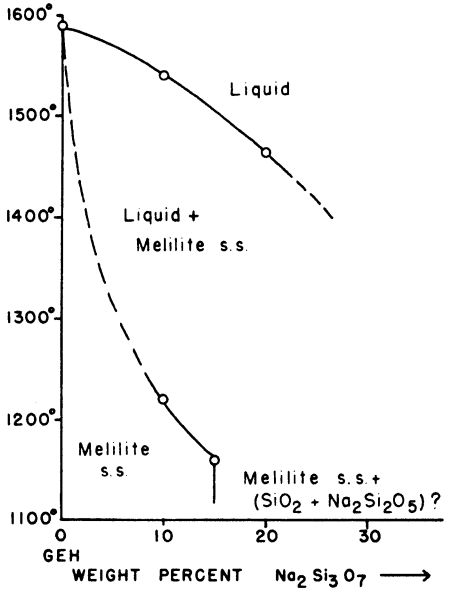



FIG. 1.—Equilibrium diagram for the system $Ca_2Al_2SiO_7$–$Na_2Si_3O_7$.

was made positively to identify the constituents present. The inhomogeneous crystallization indicates that, if any solid solution exists, it is composed of considerably less than 10 per cent by weight of $Na_2Si_3O_7$; also, the refractive indices of the akermanite were the same as those of the pure synthetic crystals ($\omega$ = 1.633; $\epsilon$ = 1.639). The fact that the indices were not changed is very strong evidence that no solid solution was formed at all.

The conclusion that akermanite takes no $Na_2Si_3O_7$ in solid solution is not sur-

prising; in fact, it is to be expected. If the formulae be written

$Ca_2MgSi_2O_7$ (akermanite)
$Na_2Si\ Si_2O_7$,

it is seen that to form mix-crystals a substitution of Si for Mg would be required. The improbability of this substitution has already been mentioned; not only is the Mg ion considerably larger than the Si ion, but there is a valence difference of two charges. This situation does not exist in the case of gehlenite:

$Ca_2Al_2SiO_7$ (gehlenite)
$Na_2Si_2SiO_7$.

The Na-Ca and subsequent Si-Al substitution is well known in many minerals.

The solid solution relations are thus such that one end-member of a completely isomorphous series accepts a limited amount of a third ingredient, whereas the other end-member does not. Mixtures of the three components might then be expected to show a drop in the acceptance of the $Na_2Si_3O_7$ as the composition changes in the direction of enrichment in akermanite.

A mix composed of 45 per cent gehlenite, 45 per cent akermanite, and 10 per cent $Na_2Si_3O_7$ was prepared. The completely devitrified preparation consisted of melilite crystals plus a few per cent of inhomogeneous material, showing that less than 10 per cent $Na_2Si_3O_7$ is taken into solid solution at the mid-point of the gehlenite-akermanite series. The optical properties of the melilite, in the preparation described above, confirm this observation; the indices of crystals composed of equal amounts of gehlenite and akermanite are $\omega = 1.653$ and $\epsilon = 1.652$ (Ferguson, Buddington, 1920), the indices of the melilite formed from the 45-45-10 per cent mixture are $\omega = 1.640$, $\epsilon$ approx. $= 1.637$. Figure 2 shows that in gehlenite, 10 per cent $Na_2Si_3O_7$ causes a drop in mean index of approximately 0.020. The drop in the above solid solution is but 0.013—additional evidence that less than 10 per cent of the soda "molecule" enters this composition.

If the amount of $Na_2Si_3O_7$ varies in a linear fashion with the akermanite-gehlenite ratio, the mid-point in the isomorphous series should contain 7½ per cent $Na_2Si_3O_7$. The amount of residual material in the one composition exam-

TABLE 1

THERMAL DATA FOR THE SYSTEM $Ca_2Al_2SiO_7$–$Na_2Si_3O_7$

| Weight Per Cent | | Time (Minutes) | Temperature (° C) | Final Condition |
|---|---|---|---|---|
| $Ca_2Al_2SiO_7$ | $Na_2Si_3O_7$ | | | |
| 90 | 10 | 30 | 1,543 | All glass |
| | | 30 | 1,539 | Sparse melilite |
| | | 120 | 1,224 | Melilite, traces of interstitial glass |
| | | 120 | 1,216 | All melilite |
| 85 | 15 | 150 | 1,165 | Melilite, traces of glass |
| | | 120 | 1,155 | All melilite |
| 82½ | 17½ | ........ | ........ | Nonhomogeneous crystallizations at subsolidus temperatures |
| 80 | 20 | ........ | ........ | Nonhomogeneous crystallizations at subsolidus temperatures |
| | | 30 | 1,464 | Sparse melilite |
| | | 30 | 1,460 | All glass |

ined, as well as the refractive index of the melilite, agrees well with this value of $7\frac{1}{2}$ per cent. Figure 3 is constructed from the above data and shows the limited extent to which soda can enter the structure of the synthetic melilites. Additional points, to fix with more certainty the position of the boundary delimiting the soda-bearing melilites, are not necessary. It is apparent that the soda solid solution is at a maximum in pure gehlenite and that the amount of soda in the synthetic melilites drops off in a linear fashion as the akermanite content of the mix-crystals increases, until reaching zero in pure akermanite.

Soda entering the melilite structure must do so at the expense of Ca, and, in so doing, either the Mg, the Al, or the Si must be replaced by an ion of higher charge. Thus, if a sodium ion replaces a calcium ion, one of the following associated anion replacements is possible; $Si^{4+}$ for $Al^{3+}$, $P^{5+}$ for $Si^{4+}$, or $Al^{3+}$ ($Ga^{3+}$) for $Mg^{2+}$. In addition to the anion substitutions, cation substitutions of a different nature must be considered. If Na replaces Ca, electroneutrality can also be restored by the replacement of oxygen by $OH^-$ or $F^-$. The Si for Al replacement is the one that takes place in the solid solution of $Na_2Si_3O_7$ in gehlenite but cannot do so in akermanite. Phosphorous analyses have not been made on natural melilites; but, if phosphorous is present, it is not at all likely that it exists in appreciable amounts; therefore, it cannot be responsible for the presence of any significant amount of soda in the structure. The third possible anion replacement, that of Al for Mg, does not take place, as is shown by the results obtained with the mix of the composition Ak 45 per cent–Geh 45 per cent–$Na_2Si_3O_7$ 10 per cent. The Al for Mg replacement in akermanite would result in the formula $NaCaAlSi_2O_7$, which is nothing more than a composite of gehlenite and $Na_2Si_3O_7$. Inasmuch as the three-"component" mix, described above, contains this "molecule" as well as akermanite and yet shows no tendency to form a solid solution over and above that formed in the gehlenite, it is obvious that Al does not replace Mg in akermanite.

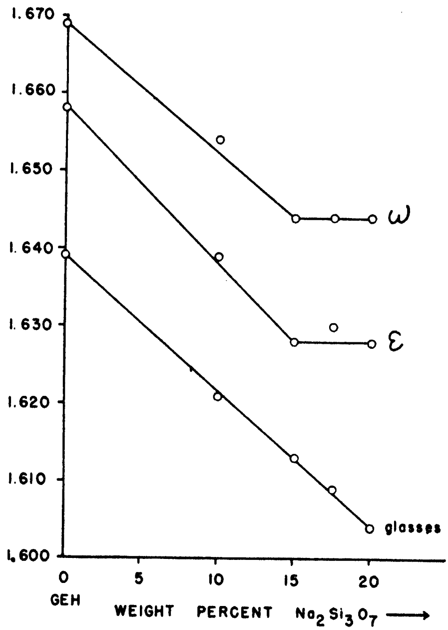



FIG. 2.—Refractive indices of melilites and glasses.

The importance of $OH^-$ in replacing $O^=$ with a resulting $Na^+$ for $Ca^{++}$ replacement in melilites cannot be evaluated with the analytical data on hand. It has been observed that in $Si_2O_7$ structures $OH^-$ does not replace $O^=$. This is an empirical fact and has no theoretical basis. On this evidence alone one might say that melilites are not likely to show this substitution. Some of the published analyses show an appreciable water content (in one case up to 1.59 per cent), but

in most cases $H_2O$ was not determined. Much of the water found in ordinary analyses probably is not present as the hydroxyl; some may come from impurities and some may be adsorbed. The ordinary means of determining $H_2O$ is not satisfactory, as it does not detect all the $OH^-$ that may exist in the crystal; it is necessary to break down the structure that soda can enter the melilites by any means other than as a replacement for Ca with the accompanying $Si^{4+}$ for $Al^{3+}$ substitution. This is tantamount to saying that the $Na_2Si_3O_7$ "molecule" is the only sodic "molecule" that can act as a substitutive end-member. The fact that the experimental results show that only 3.85 per cent $Na_2O$ can be taken up by

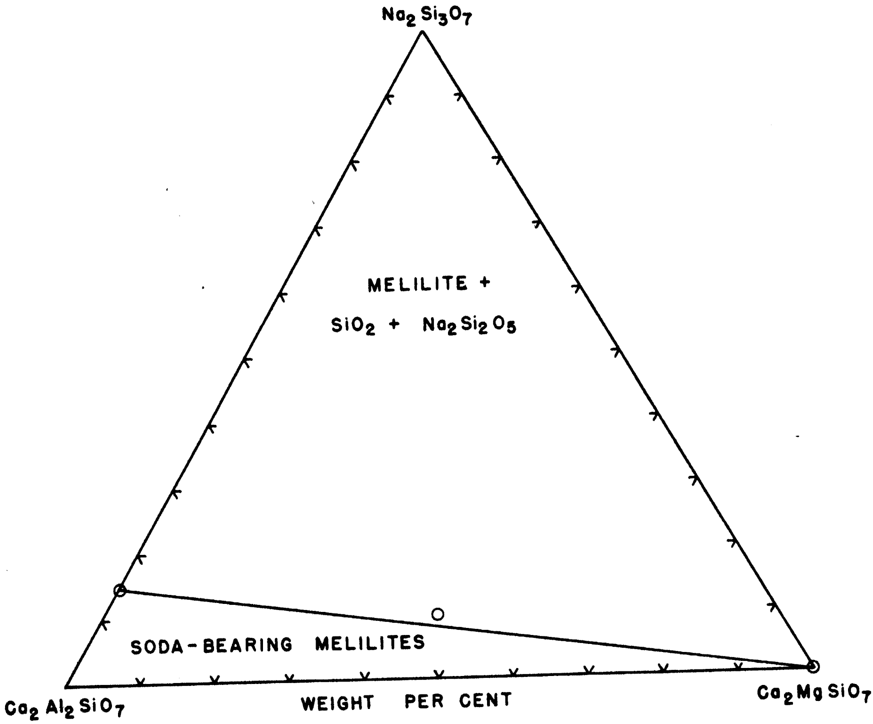



FIG. 3.—The stability field of the soda-bearing synthetic melilites

completely (for example, see Sahama, 1946) before all the hydroxyl is liberated as water. One per cent of water, present as $OH^-$ replacing oxygen, can account, however, for approximately 3 per cent of soda, and water, therefore, should be carefully determined in the course of any analytical work that is concerned with crystal chemistry.

With the possible exception of this cation replacement, it is thus improbable pure gehlenite and that the soda content drops as the akermanite content of the crystals increases makes it difficult to account for the high values of $Na_2O$ reported in natural melilites if water is not considered. It must also be remembered that, at the rather high temperatures reached in the laboratory investigation, it is likely that more solid solution takes place than is probable in nature.

The problem of the soda content of

natural melilites becomes more confusing when the chemical analyses are studied. Berman (1929) states that "the gehlenites in general have a low percentage of the minor molecules $Na_2Si_3O_7$ and $CaSi_3O_7$." Figure 4 is a plot of soda (+ potash) against magnesia (+ ferrous iron). The abscissa represents the gehlenite-akermanite ratio in simple (synthetic) melilites, as pure gehlenite contains no MgO and pure akermanite contains 14.79 per cent by weight. The previously discussed limit of soda in this system is represented by the line from 3.85 per cent at zero MgO to zero per cent at 14.79 per cent MgO. Twenty-three analyses collected by Berman (1929) are plotted, plus the $Na_2O$–MgO ratios of two analyses done since 1929 (Tilley, 1929; Schaller, 1942). Six melilites have less than ½ per cent soda, and five of these are rich in the gehlenite "molecule." The sixth is virtually pure akermanite and would be expected to show little or no $Na_2O$. Indeed, there is shown a weak tendency for the more akermanitic melilites to be the richest in $Na_2O$, which is just the opposite of what would be expected on the basis of the experimental and theoretical considerations. Figure 4 clearly illustrates the soda problem in the melilites. The $Na_2O$ that can be accounted for as $Na_2Si_3O_7$ varies from 3.85 per cent in gehlenite to zero per cent in akermanite; seventeen of the twenty-five points fall above the line delimiting this soda content.

Berman (1929) pointed out the fact that the Ca + Na content is generally below the value of 20 expected on the basis of 70 oxygen atoms. It was because of this deficiency that he proposed $CaSi_3O_7$ as an end-member. Inasmuch as Na must replace Ca and because $CaSi_3O_7$ (or any calcium silicate for that matter) does not form a solid solution with the melilites, the value of 20 atoms of Ca + Na would be expected to be constant. Figure 5 is a plot of Ca against Na (+ K). The units are atomic composition, on the basis of 70 oxygen atoms in the melilite formula. Replacement of Ca by Na is on a 1:1 basis, so that perfect substitution would result in the straight line with the negative slope of 1 shown on the diagram. The 23 analyses considered by Berman are here plotted; the tendency for the Na + Ca values to be less than 20

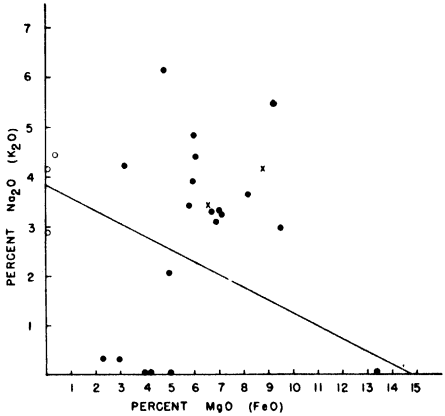



FIG. 4.—$Na_2O$(+ $K_2O$) versus MgO(+ FeO), in weight per cent, in natural melilites. Crosses are analyses since 1929. Open circles are called "sarcolites."

is obvious—only 4 points lie above the line and 4 lie on it.

The fundamental problems of the melilites have not been solved by this investigation, but it is hoped that some of the points have been clarified. The deficiency in total Ca + Na cannot be readily explained; the soda in excess of that shown to be taken up by gehlenite cannot be explained here. Berman showed that the $Y3$ member of his melilite general formula (i.e., Si, Mg, Al) is quite constant, varying but slightly from the theoretical value of 30 atoms per 70 oxygen atoms. Poor chemical analyses (most of these are quite old) might be blamed for the discrepancies except for this con-

stancy of Mg, Si, and Al. The same can be said for inclusions or impurities in the samples. The influence of pressure is unknown, but its effect must be small, as none of the melilites examined were formed under conditions of very high pressure. The influence of hydroyxl is unknown but probably not large, as already discussed. It should also be noted that if $OH^-$ substitutes for $O^=$, the soda substitution then occurs with no effect on the Si content. The fact that the melilites tend to show high Si contents might be used as an argument against the hydroxyl substitution and might favor the argument for $Na_2Si_3O_7$ solid solution.

More information is necessary on the natural minerals; careful chemical and X-ray work should be done. An X-ray investigation of the synthetic series may also prove fruitful. The possibility exists that there may be significant differences between the metamorphic and the igneous melilites, a problem that offers possibilities of profitable study with relation to other minerals as well.

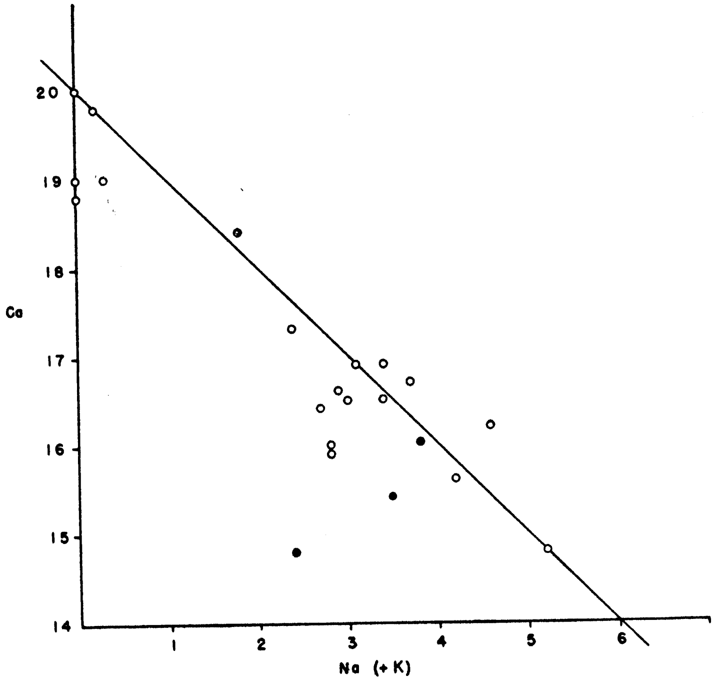



FIG. 5.—Atomic ratios of Ca versus Na(+ K) in natural melilites. Solid circles represent "sarcolites"

THE POSSIBILITY OF A CALCIUM-ALUMINATE SOLID SOLUTION IN GEHLENITE

In a recent paper dealing with a portion of the $Na_2O–CaO–Al_2O_3–SiO_2$ system (Goldsmith, 1947) the possibility of a solid solution of $CaO \cdot 2Al_2O_3$ in gehlenite was considered. The improbability of obtaining mix-crystals of gehlenite and the aluminate was pointed

out, as a substitution of the small Al ion for the large Ca ion is involved.

Ca Ca $AlAlSiO_7$ (gehlenite)
Ca Al $AlAlAlO_7$ ($CaO \cdot 2Al_2O_3$)

The substitution of Al for Ca, as far as the writer knows, is not recognized in any known structure.

It was felt that the problem was interesting enough to warrant an investigation, and, accordingly, glasses were prepared in the gehlenite-rich portion of the system $Ca_2Al_2SiO_7$–$CaO \cdot 2Al_2O_3$. The glasses were devitrified at approximately 1,300° C. and examined microscopically. Mixtures up to 30 per cent $CaO \cdot 2Al_2O_3$ were optically homogeneous, i.e., but a single phase could be detected. The material devitrified as a somewhat fibrous, finely crystalline product with parallel and subparallel crystallographic orientation, as is common when a glass is crystallized well below its solidus temperature. A rather small lowering of the refractive indices was noted; the 30 per cent $CaO \cdot 2Al_2O_3$ mix showed $N = 1.658$, $n = 1.645$, a drop of approximately 0.01 from the indices of pure gehlenite.

No solid solution between gehlenite and any calcium aluminate was found in the original work on the system $CaO$–$Al_2O_3$–$SiO_2$ (G. A. Rankin and F. E. Wright, 1915). The eutectic between gehlenite and $CaO \cdot 2Al_2O_3$[2] is located at 31 per cent $CaO \cdot 2Al_2O_3$, and it appeared to the writer that even if the gehlenite–$CaO \cdot 2Al_2O_3$ solid solution existed, it is improbable that it would extend all the way to the eutectic. In the earlier stages of this work, on compositions low in $CaO \cdot 2Al_2O_3$, the apparent optical homogeneity led to the conclusion that solid solution between gehlenite and the calcium aluminate certainly took place. It was on this basis that an abstract was published,[3] but additional work has cast considerable doubt on the validity of the above conclusion.

The suspicion that the optical data were perhaps misleading led to an investigation of the devitrified products by means of X-ray diffraction. Powder pictures were made of pure gehlenite, pure $CaO \cdot 2Al_2O_3$, and of a mechanical mixture of 90 per cent gehlenite and 10 per cent $CaO \cdot 2Al_2O_3$. It was found that considerably less than 10 per cent $CaO \cdot 2Al_2O_3$ can be detected in gehlenite in X-ray powder pictures, because the three strongest calcium-aluminate lines were easily detected in the picture of the composite preparation. Powder pictures were then made of the mixtures that had been devitrified at approximately 1,300° C., with the result that the 90 per cent gehlenite–10 per cent $CaO \cdot 2Al_2O_3$ showed none of the lines of the $CaO \cdot 2Al_2O_3$, whereas those of 20 per cent and higher $CaO \cdot 2Al_2O_3$ content did. These data revealed that microscopic examination was insufficient in this system. A submicroscopic intergrowth of gehlenite and $CaO \cdot 2Al_2O_3$ apparently formed when the glasses were devitrified at 1,300° C., which appeared to be optically homogeneous. It also appeared, however, that some solid solution was indeed taking place, as the 10 per cent mixture did not show calcium-aluminate lines. If any line shifts in the powder pictures are present, they are so small as to be within the limits of error of measurement.

The liquidus relations (originally determined with $3CaO \cdot 5Al_2O_3$ as one end-member by Rankin and Wright [1915]) were not reinvestigated because of the high temperatures involved. Because the X-ray evidence on the 10 per

[2] Erroneously identified as $3CaO \cdot 5Al_2O_3$ in the original investigation and subsequently labeled as such in all later publications and diagrams in this country.

[3] Geol. Soc. America Bull. 58, no. 12 (pt. 2), December, 1947, p. 1183.

cent $CaO \cdot 2Al_2O_3$ mix indicated some solid solution, it was thought worth while to determine the solidus curve. In the attempt to locate the solidus, using both previously devitrified material and glass, all mixtures showed marked inhomogeneity when held at high temperatures. The 10 per cent mix, for example, which was homogeneous at 1,300° C., showed numerous tiny blebs and inhomogeneities at 1,500° C., and the X-ray picture indicated the presence of $CaO \cdot 2Al_2O_3$. The phenomenon of homogeneity and apparent solid solution at 1,300° C., with inhomogeneity at higher temperatures, is indeed unusual, the opposite of what normally would be expected. The X-ray evidence of the presence of $CaO \cdot 2Al_2O_3$ in mixtures with more than 10 per cent of the component $CaO \cdot 2Al_2O_3$ cannot thus be explained by unmixing at low temperatures from a high-temperature solid solution, as it is obvious that little or no solid solution exists at near-solidus temperatures. A solid solution stable at low temperature and unstable at higher temperatures is difficult to explain unless one assumes an inversion in gehlenite itself; no evidence for such an inversion has ever been noted.

Intimate mixtures of very finely ground crystalline gehlenite and $CaO \cdot 2Al_2O_3$ were heated in an attempt to produce reaction in the solid state. If solid solutions were formed between these components, it might be expected that the end-members would react in the solid state to produce mix-crystals. Heating for 3 days at various temperatures up to 1,500° C. produced no change in microscopic appearance or X-ray diffraction pattern; the two phases could easily be detected (90 per cent gehlenite, 10 per cent $CaO \cdot 2Al_2O_3$). However, these results do not conclusively prove that solid solution cannot take place. It is not improbable that a solid-state reaction of this sort might not always take place (or do so very slowly), particularly if the solid solution is a rather unstable one. It is not unlikely that mix-crystals might be obtained by direct crystallization from a melt (glass) in which but a very small rearrangement of atomic positions is necessary; but if the solid solution is a relatively unstable one, the necessary amount of rearrangement might not take place from the two stable crystalline phases. To do so would require the destruction of the lattices of the two crystalline phases, with relatively large-scale migration and construction of a new single phase.

Experimental work performed on the subsolidus relations between $Ca_2Al_2SiO_7$ and $CaO \cdot 2Al_2O_3$ thus shows rather conflicting evidence on the question of solid solution. Devitrification of glasses at approximately 1,300° C. apparently results in a single phase, whereas at 1,500° C. an obviously inhomogeneous product results. It appears that at 1,300° C. a solid solution with an upper limit of approximately 10 per cent $CaO \cdot 2Al_2O_3$ may form, although several other observations tend to contradict this conclusion. No line shifts or intensity changes could be detected in the X-ray powder pictures, no observable reaction in the solid state takes place in 3 days' time, and little or no faith can be put in optical examination of this system.

If there is solid solution of a calcium aluminate in gehlenite, it is almost certainly not present in the rock-forming minerals. The melilites are not undersilicated but generally tend to show excess silica, as mentioned earlier. It has also been stated that the Y term of Berman (Si + Mg + Al) is constant at a value of 30, and such would not be the case if $CaO \cdot 2Al_2O_3$ were present in the crystals.

The intergrowths of $CaO \cdot 2Al_2O_3$ in gehlenite that could not be resolved opti-

cally are interesting if one considers that such intergrowths may be present but unrecognized in certain natural minerals. The possibility of fine intergrowths giving misleading information on chemically analyzed minerals should be considered.

## SUMMARY AND CONCLUSIONS

The solid solution of $Na_2Si_3O_7$ in gehlenite ($Ca_2Al_2SiO_7$) was experimentally determined and takes place to the extent of 15 per cent by weight (3.85 per cent as $Na_2O$). Soda cannot enter the structure of pure akermanite, and the amount of $Na_2O$ that can enter mix-crystals of synthetic gehlenite and akermanite varies in a linear fashion from 3.85 per cent in gehlenite to 0 in akermanite. The high soda content found in some natural melilites (up to approximately 6 per cent) cannot be accounted for, on the basis of this investigation, by solid solution of $Na_2Si_3O_7$ in the melilites. It is shown that the only reasonable way that soda can enter the melilite structure, other than as the "end-member" $Na_2Si_3O_7$, is by $OH^-$ replacement of $O^=$; the magnitude of this replacement has not been fully evaluated. The phase diagrams of gehlenite-wollastonite and akermanite-wollastonite show no solid solution complications and are truly binary in nature, indicating that the melilites do not take any calcium silicate (of any formulation) in solid solution. This fact, plus structural considerations, eliminates any of the earlier proposed lime end-members, such as Berman's "submelilite," $CaSi_3O_7$. Additional careful chemical (including $OH^-$ and $F^-$ determinations) and X-ray work on natural melilites is necessary.

The possibility of the solid solution of $CaO \cdot 2Al_2O_3$ in gehlenite is considered. There is some indication that approximately 10 per cent $CaO \cdot 2Al_2O_3$ may be taken up by gehlenite, but the experimental evidence is somewhat contradictory. If such a solid solution does exist, it is indeed unusual in that the substitution of Al for Ca is involved.

ACKNOWLEDGMENTS.—This work was done in conjunction with a research program made possible by financial assistance from the Office of Naval Research. The author is also indebted to Professors Tom. F. W. Barth and F. E. Wickman for stimulating discussions and suggestions, and to Mrs. Ursula Chaisson, who took all the X-ray photographs used.

## REFERENCES CITED

BERMAN, HARRY (1929) Composition of the melilite group: Am. Mineralogist, vol. 14, pp. 389–407.

BOWEN, N. L.; SCHAIRER, J. F.; and POSNJAK, E. (1933) The system, CaO–FeO–$SiO_2$: Am. Jour. Sci., vol. 26, pp. 193–284.

BUDDINGTON, A. F. (1922) On some natural and synthetic melilites: Am. Jour. Sci., vol. 3, pp. 35–87.

FERGUSON, J. B., and BUDDINGTON, A. F. (1920) The binary system akermanite-gehlenite: Am. Jour. Sci., 4th ser., vol. 50, pp. 131–140.

GOLDSMITH, J. R. (1947) The system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$: Jour. Geology, vol. 55, pp. 381–404.

KRACEK, F. C. (1930) The system sodium oxide–silica: Jour. Physical Chemistry, vol. 34, pp. 1583–1598.

MACHATSCHKI, FELIX (1930) Die Summenformel der melilithe: Centralbl. Mineralogie, Geologie, und Paläontologie, pp. 278–284.

OSBORN, E. F., and SCHAIRER, J. F. (1941) The ternary system pseudo-wollastonite-akermanite-gehlenite: Am. Jour. Sci., vol. 239, pp. 715–763.

RANKIN, G. A., and WRIGHT, F. E. (1915) The ternary system of CaO–$Al_2O_3$–$SiO_2$: Am. Jour. Sci., 4th ser., vol. 39, pp. 1–79.

SAHAMA, TH. G. (1946) On the chemistry of the mineral titanite: Comm. géol. Finlande Bull. 19, pp. 88–120.

SCHALLER, W. T. (1916) Mineralogic notes: Nova Scotia Geol. Survey Bull. 610, 3d ser., pp. 109–128.

——— (1942) Analysis of coarse melilite (from uncompahgrite, Iron Hill stock): U.S. Geol. Survey Prof. Paper 197-A.

TILLEY, C. E. (1929) On melilite as a product of interaction of limestone and basaltic liquid (analysis by H. F. Harwood): London Geol. Mag., vol. 66, p. 347.

WARREN, B. E. (1930) The structure of melilite, $(Ca,Na)_2(Mg,Al)(Si,Al)_2O_7$: Zeitschr. Kristallographie, vol. 74, pp. 131–138.

WINCHELL, A. N. (1924) The composition of melilite: Am. Jour. Sci., 5th ser., vol. 8, pp. 375–384.

# RADIAL DIFFUSION AND CHEMICAL STABILITY IN THE GRAVITATIONAL FIELD[1]

HANS RAMBERG
University of Chicago

ABSTRACT

Because gravity influences chemical equilibrium, consideration of the thermodynamics of our globe requires understanding of chemical processes and stability in the gravitational field. The influence of gravity on stability of mixed condensates (crystals or magmas) and of stoichiometric crystals is analyzed. The theoretical analysis supports the hypothesis of granitization and metasomatism by upward diffusion of some elements.

## INTRODUCTION

In recent years an increasing number of students of granitization have realized that these phenomena cannot be completely accounted for without assuming large-scale diffusion through the solid rocks (Wegmann, 1935; Roubault and Perrin, 1939; Backlund, 1946; Barth, 1948*b;* Ramberg, 1944*a*, *b*, 1945).

In the earth the principal laws governing radial diffusion have never, so far as I know, been fully understood by geologists. Some years ago the writer (1944*a*, 1945, 1946) proved that mechanical instability in the gravitational field results in chemical or thermodynamic instability. In other words, if rocks or magmas with comparatively high density rest on rocks or magmas with low density, then chemical potential differences exist, tending to make the two bodies change places with the help of dispersion, vertical diffusion, and consolidation.

This paper is devoted to the principal laws of radial diffusion in the earth. We must base our estimate of the role of diffusion in the metasomatic alteration of the earth's crust in part on these principles.

[1] Presented at the University of Oslo, April, 1946, as the third part of a series on the "Thermodynamics of the Earth's Crust." The first two parts have already been published (Ramberg, 1944*b* and 1945). Manuscript received December 15, 1947.

In the earth's crust and even deep in the earth, most material is condensed and contains only negligible quantities of free gases. Therefore, chemical stability in the earth can be said to be equivalent to equilibria of diffusion. Such equilibria are not characterized by diffusion of matter in the form of activated molecular particles (atoms, ions) in but one direction, even if infinite intervals of time are considered. The equilibrium is believed to be dynamic, that is, equal numbers of activated molecular particles migrate in opposite directions through any given cross section in the earth.

Like other chemical processes, the diffusion currents are driven by chemical forces which can be expressed in terms of the chemical potential, $\mu$; free energy, $\zeta$; activity, *a;* fugacity, *f;* and vapor tension, $\pi$. These terms are not identical but are related by means of more or less simple equations.

As pointed out in previous papers (1944*a*, 1946), the writer finds it most convenient to make use of the term "partial vapor tension," $\pi$, if both diffusion and chemical equilibrium are involved and even if the system is completely condensed. The term "partial vapor tension" is accordingly considered an important thermodynamic property of condensates,

being useful whether a vapor is formed or not.

Because some geologists (I. Oftedal, 1947) have criticized the concept of vapor tension of minerals as indicative of the chemical potential, the views of the distinguished physicochemists, Lewis and Randall, are significant (Marsh, 1935, p. 31):

> One of the most striking results of this character is obtained if we calculate the vapor pressure of tungsten at 25° from experiments at very high temperatures. The result, $10^{-149}$ atmospheres, would mean that the concentration of tungsten vapor would be less than one molecule in a space equivalent to the known sidereal universe. Such a calculation need not alarm us. Allowing for the possibilities of experimental uncertainties, we may use such a calculated vapor pressure in our thermodynamic work with the same sense of security as we use the vapor pressure of water.

Even in crystalline systems, diffusion resistance is never infinite (Barrer, 1941), e.g., horizontal chemical potential or vapor-tension gradients will always give rise to diffusion of activated particles, although the rate of this process in some cases will be very slow (Jost, 1937). Chemical equilibrium in a horizontal plane in the earth's crust is therefore, as we know, characterized by no gradient of the partial chemical potentials.

Equations of the following type:

$$\frac{dm}{dt} = -qD'\frac{d\pi}{dx} \qquad (1)$$

hold for the diffusional transfer and along a horizontal plane in the earth. The quantity $dm/dt$ is the power of diffusion through section $q$; $D'$ is a diffusion coefficient depending on the character of the system, on $P$ and $T$, and on the type of migrating particles. In heterogeneous systems such as rocks, $D'$ is to be considered a mean value depending on the diffusion coefficients for the intergranular diffusion, the mosaic-fissure diffusion, and the volume diffusion in the different phases (minerals, pore solutions). The quantity $d\pi/dx$ is the partial tension gradient in the horizontal direction.

Obviously, stability exists when $dm/dt = 0$, i.e., when $d\pi/dx = 0$, since $q$ and $D'$ are finite quantities.

Considering, now, the effect of gravity, it is clear that, in order to insure chemical stability in a vertical direction, the gravitational attraction affecting the diffusing activated particles must be compensated by a vapor-tension gradient acting in opposition to gravitation.

Chemical stability in the gravitational field is achieved if the partial vapor tension of a given element in the crystalline mantle of the earth increases with depth in accordance with the simple equation (Ramberg, 1944*a*),

$$\pi_y = \pi_0 e^{(My)/(RT)}, \qquad (2)$$

where $\pi_y$ is the partial vapor tension at depth $y$; $\pi_0$ is the stable partial tension at level 0 (e.g., the surface); $M$ is molecular weight of the particle considered; and $R$ is the gas constant. Hence, the stable partial tension gradient is

$$\frac{d\pi}{dy} = \pi\frac{M}{RT}, \qquad (3)$$

when $y$ is positive downward.

Partial tension gradients steeper than those corresponding to equation (3) make the corresponding elements diffuse upward through the crust because in this case the upwardly directed chemical force is stronger than the gravitational attraction. On the other hand, diffusion has to go downward if the partial tension gradient is less than that expressed in equation (3).

Diffusion parallel to the gravitative force-lines can accordingly be expressed by an equation of the following form:

$$\frac{dm}{dt} = -qD'\left(\frac{d\pi}{dy} - \pi\frac{M}{RT}\right). \qquad (4)$$

This equation demonstrates that $dm/dt$ is zero (i.e., chemical equilibrium exists) when $d\pi/dy = \pi(M/RT)$, and that the diffusion current is running upward if $d\pi/dy > \pi(M/RT)$, and downward if $d\pi/dy < \pi(M/RT)$.

Considering the chemical potential $\mu$, equilibrium conditions are characterized by no change in $\mu$ throughout the system (MacInnes, 1939, p. 174). This means that the relation between vapor tension and chemical potential changes with position in a gravitational field.

### THE STABILITY OF MIXED CRYSTAL AND MAGMA LAYERS IN THE GRAVITATIONAL FIELD

Let us first consider a thick layer of an ideal mixed crystal, say $A_mB_n$, in the gravitational field. For convenience, $A$ and $B$ are elements, e.g., metals. Furthermore, let the temperature be constant and the composition of the top of the layer (which does not necessarily reach up to the earth's surface) be $A_mB_n$.

According to Rault's law, the partial vapor tension of the ideal mixed crystal at the top will be

$$\pi_{A_m} = \pi_A \frac{m}{n+m} \tag{5}$$

for the partial $A$-tension, and

$$\pi_{B_n} = \pi_B \frac{n}{n+m} \tag{6}$$

for the partial $B$-tension. Here $\pi_A$ and $\pi_B$ are the vapor tensions of the metals $A$ and $B$ in their pure state; $m$ is the number of mols of $A$ in the mixed crystal; and $n$ the number of mols of $B$. If there is stability throughout the whole layer, the partial tensions at every depth, $y$, must coincide with the law of stability (eq. [2]) viz:

$$\pi_{A_y} = \pi_A \frac{m}{n+m} e^{M_A y/RT} \tag{7}$$

and

$$\pi_{B_y} = \pi_B \frac{n}{n+m} e^{M_B y/RT}, \tag{8}$$

where $\pi_{A_y}$ and $\pi_{B_y}$ are the stable partial tensions with reference to $A$ and $B$ in the depth $y$; and $M_A$ and $M_B$ are the atomic weights of $A$ and $B$, respectively.

The pressure gradient will make the partial tension increase downward, but only in the case of monatomic condensates will this downwardly increasing vapor tension harmonize with the law of chemical stability in the gravitational field (eq. [2]). In mixed crystals there will be a continuous change in composition.

The mechanical pressure at depth $y$ is $P_y = d \cdot y$ where, for convenience, the density $d$ is supposed to be independent of both pressure and change in composition of the mixed crystal layer.

According to the equation expressing the relation between vapor tension and mechanical pressure, the pressure in depth $y$ makes the partial vapor tension rise in the following manner, provided that composition is constant:

$$\begin{aligned}\pi_{A_P} &= \pi_A \frac{m}{n+m} e^{P_y V_A/RT} \\ &= \pi_A \frac{m}{n+m} e^{d \cdot y V_A/RT}\end{aligned} \tag{9}$$

and

$$\begin{aligned}\pi_{B_P} &= \pi_B \frac{n}{n+m} e^{P_y V_B/RT} \\ &= \pi_B \frac{n}{n+m} e^{d \cdot y V_B/RT},\end{aligned} \tag{10}$$

respectively. Here $V_A$ and $V_B$ are the fictive molal volumes of the components $A$ and $B$, respectively, in the mixed crystal. (The "fictive molal volume" of a given component in a mixture is the change in volume that an infinitely large body of the mixture undergoes by reversible loss or gain of 1 mol of the considered component.) The fictive molal

volume of a certain ion or atom is different in different phases. Likewise, it changes with changing composition and physical condition of a given phase.

It follows that if equation (7) is identical with equation (9) and equation (8) is identical with equation (10), then the stable conditions correspond to a homogeneous mixed crystal layer. On the other hand, if these equalities do not exist, i.e., if,

$$\pi_{A_P} \gtrless \pi_{A_y}, \quad \text{or} \quad d \gtrless \frac{M_A}{V_A}, \quad (11)$$

and

$$\pi_{B_P} \gtrless \pi_{B_P}, \quad \text{or} \quad d \lesseqgtr \frac{M_B}{V_B}, \quad (12)$$

then constant composition of the layer will result in a chemically unstable state.

If $d > M_A/V_A$ and $d < M_B/V_B$, there will exist forces which make the element $A$ migrate upward and $B$ downward. If $d < M_A/V_A$ and $d > M_B/V_B$, then $A$ migrates down and $B$ up. The change in composition at different levels thus engendered leads to the following conditions: $\pi_{A_P} = \pi_{A_y}$ and $\pi_{B_P} = \pi_{B_y}$, obviously due to the alteration of the term $n/(n + m)$ and $m/(n + m)$ in equations (9) and (10). Thus we find the stable composition, $(m/[n + m])_y$, at depth $y$ when the surface composition is $m/[n + m]_0$

$$\left(\frac{m}{n+m}\right)_y = \left(\frac{m}{n+m}\right)_0 e^{(M_A y - d \cdot y V_A)/RT} \quad (13)$$

(see also MacInnes, 1939, p. 175). The term $M/V$ thus governs the change in composition with depth in ideal mixed crystals. I have named this quantity the "fictive density" of the particle (ion, atom) in question (Ramberg, 1946). In harmony with what is said above about the fictive molal volume, the fictive density of a certain element or ion changes from one crystal to another. If the fictive densities of elements in a substitution mixed crystal of type $A_mB_n$ differ from the density of the crystal itself, then the forces of differentiation will try to change the composition in a vertical direction in the gravitational field. Elements with fictive densities larger than the density of the crystal will be gradually concentrated at lower levels and vice versa.

In silicate minerals or melts these principles may be applied to elements that mutually replace each other. Because such elements have nearly equal fictive molal volume, their fictive densities are approximately proportional to the atomic weight. Consequently, in a number of ferromagnesian minerals and in mixed crystals of the plagioclase types, the stable condition is characterized by a gradual downward increase in the Fe/Mg ratio and the Ca/Na ratio, respectively. Provided that the fictive molal volumes of the mutually replacing elements are equal, equation (13) leads into:

$$\left(\frac{m}{n}\right)_y = \left(\frac{m}{n}\right)_0 e^{[(M_A - M_B)y]/RT}. \quad (14)$$

This equation can be used with fair accuracy for calculating the change in the Ca/Na ratio with depth in the plagioclase mixed crystal series. Assume, for example, a temperature of 400° K. and Ca/Na = 1 at the surface, there is, at a depth $y$ = 20 km., a stable ratio of Ca/Na = 2.7. In ferromagnesian minerals a change of Fe/Mg ratio from 1 to 2.7 extends from the top to about 9 km. depth.

In addition-mixed crystals, some elements may enter vacant places in the lattice and give rise to only a small expansion of the unit cell, i.e., the fictive molal volume may in this case be very small (in very rare cases it may be zero

or even negative). This means that the fictive density of the elements that occupy the vacant places may be very large, even if the elements have a small atomic weight. At stable conditions the incorporated elements in this case will be concentrated at lower levels, in spite of being incorporated in a dense crystal.

In mixed crystals of metal and hydrogen the unit cell does not necessarily increase through the absorption of hydrogen because the small H-atoms find their place in the interstices among the metal atoms. In some cases even, contraction may take place, resulting in a negative fictive molal volume of hydrogen so that the $H_2$-tension of the hydrogen-metal mixture decreases with increasing external pressure (eq. [9]). Therefore, if hydrogen-metal mixtures exist in the earth, the very light hydrogen atoms will be concentrated at lower levels. The degree of concentration, however, is obviously limited by the miscibility between metal and hydrogen.

These peculiarities must be taken into consideration when speculations as to the condition of our globe's deeper parts are discussed (Kuhn and Rittmann, 1941).

### THE GRAVITATIVE STABILITY OF STOICHIOMETRIC CRYSTALS

In this type of crystal there are very narrow fields of homogeneity, so that the element ratio cannot diverge much from the stoichiometric ratio. On the other hand, even very negligible change in composition gives rise to considerable change in partial vapor tension. In a system of elements which are able to combine into different stoichiometric solid compounds, gravitation will obviously tend to develop a mechanically stable arrangement of different layers with different compositions and different densities. This process will take place with the help of bodily up-and-down motion of rocks and magmas in the earth.

It has, however, never been fully recognized by geologists that such a mechanically stable arrangement in the earth also corresponds to thermodynamic stability. This means that any large-scale structure of the earth not corresponding to a stable concentric-layer arrangement possesses potential chemical forces tending to develop the stable arrangement. In this connection it cannot be overemphasized that the only process able to equalize chemical instabilities in the earth is diffusion.

To explain the principal features, let us consider a theoretical system consisting of the elements $A$ and $B$ which can combine into a stoichiometric phase $C = AB$. The atomic weight of $A$ is supposed to be less than that of $B$: $M_A < M_B$. The densities of $A$, $B$, and $C$ are $d_A$, $d_B$, and $d_C$, respectively, where $d_A < d_C < d_B$.

The vapor tension of $A$ is $\pi_A$ and of $B$, $\pi_B$ at $P = 0$ and a given temperature. $C$ is supposed to have a partial $A$-tension equal to $\pi_{C_A}$ and a partial $B$-tension equal to $\pi_{C_B}$ at $P = 0$. Now $\pi_{C_A} < \pi_A$, and $\pi_{C_B} < \pi_B$, so that the reaction:

$$A + B \xrightarrow{\text{High } T,\ T\mu,\ \text{Low } T} C\,, \tag{14}$$

goes to the right at zero pressure and at the temperature under consideration.

As in mixed crystals, there will exist an instability in a homogeneous thick $C$ layer in the earth if

$$\frac{M_A}{d_C} \gtrless V_{A_C} \quad \text{and} \quad \frac{M_B}{d_C} \lessgtr V_{B_C}\,,$$

where $V_{A_C}$ and $V_{B_C}$ are the fictive molal volumes of $A$ and $B$ in the compound $C$. The element with fictive molal volume larger than $M/d_C$ tends to migrate upward and vice versa.

In this example, $V_{Ac} > M_A/d_C$ and $V_{Bc} < M_B/d_C$, so that $A$ will tend to migrate toward the top and $B$ downward. These migrations in combination with phase transitions will split up the homogeneous $C$ layer, creating a stable arrangement of three layers with $A$ on the top, $C$ in the middle, and $B$ at the bottom. The thicknesses of the layers in this stable arrangement depend on the character of the system, on temperature, and on the gravitational force. At temperatures just below the transition temperature of equation (14) (at $P_2 = 0$) the stable $C$ layer will be thin, but at lower temperatures the thickness of $C$ may be very large at stable conditions.

It is also of interest to consider the case of the compound $C = (AB)$ with a density less than $d_A$ due to a loose-packed lattice of $C: (d_C < d_A < d_B)$. In this case the potential mechanical energy is at a minimum, and mechanical stability exists when $C$ exists as a top layer above $A$ and $B$ lies at the bottom. It can also be proved that this arrangement represents the true chemical stability, otherwise a circular process arises (Ramberg, 1944*a*).

If such a light $C$ layer is exposed to the differentiational effect of gravity, then dispersion, diffusion, and consolidation will develop an $A$ layer in the middle parts of the original $C$ layer, while $B$ atoms migrate farther down, consolidating as a pure $B$ layer at the bottom.

In our first example $(d_A < d_C < d_B)$, we may have originally a thick $A$ layer lying above a $B$ layer. Then reaction and diffusion across the boundary generate an intermediate $C$ layer, which eventually attains the stable thickness. The same original arrangement in case 2 (in which $d_C < d_A < d_B$) will not give rise to interaction of $A$ and $B$ along the boundary (provided that the $A$ layer is thick enough to create a sufficiently high pressure along the boundary), but $B$ migrates up through the $A$ layer, unable to form $C$ in the lower parts of the $A$ layer. Above a certain level in the $A$ layer, however, $C$ begins to generate until a stable $C$ layer is formed at the top. Similarly, if $d_C > d_B > d_A$, we find that diffusion processes tend to rearrange the system, leaving $C$ at the bottom and $A$ and $B$ above.

The system Fe–$O_2$ and their compounds, with which the writer has briefly dealt previously (Ramberg, 1946) and which Barth has tried to treat quantitatively in this journal (1948*a*), harmonizes with the above considerations.

In a sufficiently thick layer of magnetite in the earth, there exist forces tending to make oxygen migrate upward and iron downward if the fictive molal volumes of iron and oxygen satisfy the following requirement.

$$V_{O_{mt}} > \frac{M_O}{M_{Fe}} V_{Fe_{mt}}, \qquad (15)$$

where $V_{O_{mt}}$ and $V_{Fe_{mt}}$ are the fictive molal volumes of O and Fe in magnetite.

Although the fictive molal volumes may differ considerably from the so-called "ionic" or "atomic" volumes of the different elements, it seems very reasonable that, for magnetite, equation (15) is satisfied and also that $V_{O_{mt}} > V_{Fe_{mt}}$. Thus gravity tends to differentiate the magnetite layer, establishing an arrangement of $O_2$ gas on the top (the pressure of this gas corresponds to the oxidizing pressure of magnetite at the given temperature), hematite layer below, magnetite layer in the middle, FeO farther down, and iron in its pure state at the bottom.

Because magnetite actually has somewhat lower density than hematite, however, there exists a small instability in this arrangement. The true chemical and

mechanical stability exists when the magnetite and hematite layers exchange places, so that the less oxidized and less dense phase is in contact with the $O_2$ atmosphere. This may seem impossible at first glance, but the $O_2$ tension at the surface in the first-mentioned arrangement equals the oxidizing tension of the reaction $2Fe_3O_4 + \frac{1}{2}O_2 \rightleftarrows 3Fe_2O_3$ at the given $T$. Because the rearrangement from the unstable to the true stable state is combined with a certain depression of the $O_2$ tension throughout the whole system, it is clear that magnetite will not be oxidized along the boundary against the dilute $O_2$ gas. The $O_2$ tension above the magnetite layer is somewhat lower than the oxidizing tension, whereas the oxygen tension below a certain depth is, at stable conditions, able to oxidize the magnetite into a hematite phase.

The processes in the original homogeneous magnetite layer may be explained as follows: Fe will diffuse downward from the middle parts of the layer considerably more rapidly than O migrates upward, so that the iron-poor phase, hematite, is formed in the middle parts of the original layer. Oxygen is comparatively strongly bound to the hematite phase, which absorbs the O particles migrating up from the deeper levels; only a few oxygen ions or atoms are able to escape from the hematite layer and eventually reach the surface.

Whatever the pure kinetics of the process, it is clear that the ideal stable arrangement of a Fe-$O_2$ system in the gravitational field is not a homogeneous iron oxide phase but several layers arranged as explained above. However, a minimum thickness of the original homogeneous magnetite or hematite is required to start and maintain the differentiation process. Likewise, the thicknesses of the several established layers are also definite values depending on composition of the system (the mean Fe/O ratio) and on temperature.

With certain modifications, this ideal system is of interest in discussion of the degree of oxidation in the earth at different levels. Temperature, pressure, and chemical environments are not the only factors influencing the degree of oxidation (Goldschmidt, 1943); the differentiational effect of gravity plays an important role, even in crystalline and condensed systems.

This idea of chemical squeezing of some elements with low fictive density out of the deeper parts of the earth was presented some years ago (Ramberg, 1944*a*, 1946). The degree of oxidation at different levels in the crust or the ratio of oxygen to metallic ions was considered partly due to upward diffusion of oxygen and downward migration of elements with high fictive density. Later Barth (1948) calculated the ratio between oxygen and metallic atoms in rocks from different depths and arrived at a conclusion in full agreement with the present writer. It is also interesting that Harrison Brown and Claire Patterson recently (1948) arrived at similar ideas concerning the content of oxygen in meteorites and in the earth. They concluded from investigations of meteorites that oxygen might be squeezed out from deeper parts of the planets.

In the earth today the atmosphere $O_2$ tension is very much higher than the oxidizing tension of magnetite at low temperatures. It is therefore impossible to assume that $O_2$ will migrate out of the earth's surface owing to gravitative differentiation in a solid earth at the present comparatively low temperatures of the crust. Provided that temperature was sufficiently high in an earlier evolutionary stage of the earth, gravitative differ-

entiation might have contributed to the formation of an oxygen atmosphere by chemical squeezing of oxygen out of deeper levels in a molten or solid earth. In more recent times similar processes will take place at greater depths, where temperature is high and the effect of the atmospheric oxygen does not disturb the processes.

Other examples of the influence of gravitation on chemical stability are the hydrous minerals, such as micas, hornblende, and others. Gravitation makes such minerals chemically unstable below certain depths, owing to the chemical squeezing of $H_2O$ out of the lattices and the upward migration of the water or independent $H^+$ and $O^{--}$ ions.

If hydrous minerals, e.g., hornblende, biotite, or muscovite, exist below a certain level (which depends on temperature), then the superincumbent load raises the partial $H_2O$ tension (owing to the large fictive molal volume of $OH^-$ in the lattice) above that which corresponds to vertical diffusional stability. Consequently, forces exist that make $H_2O$ (or $H^+ + O^{--}$) migrate toward the top, eventually resulting in hydrothermal metasomatism. Below a certain depth, which seems to be impossible to calculate because of lack of data, the chemical squeezing of water is so intense that an hydrous minerals like pyroxene, garnets, etc., form at the cost of mica and hornblende.

In micas a chemical squeezing of the large potassium ions will take place; the potassium tends to migrate upward, to a small extent contributing to the granitization process.

This process is of interest for the formation of the *granulite facies* of rocks. The chemical squeezing-out of the water of the hydrous minerals at high pressure and great depths indicates that high temperature is not necessary for the formation of the anhydrous minerals of the granulite facies.

The existence of olivine-rich rocks at great depths may be explained to some extent in a similar way. At a certain depth, Si and $O_2$ will be squeezed out of the hypersthene lattice. Upward migration of these elements will take place, resulting in Si-metasomatism at higher levels and olivine formation at lower levels.

Finally, let us consider the differential effect of gravitation on the well-known reaction:

$$\text{Calcite} + \text{quartz} \rightleftarrows \text{wollastonite} + CO_2\,. \tag{16}$$

At the low $CO_2$ tension existing in the air, wollastonite is formed at about 500°. In closed systems where $CO_2$ is exposed to the total rock pressure the reaction temperature rises with increasing pressure, according to V. M. Goldschmidt (1912).

The wollastonite reaction may also be considered under those conditions where the $CO_2$ gas is able to migrate away from the place of reaction. In this case the reaction temperature depends not only on rock pressure but also on the partial $CO_2$ pressure at the given place. The particular case in which the partial pressure of $CO_2$ throughout the crust is equal to the $CO_2$ tension of the atmosphere is treated in another paper (in press). In this case the reaction temperature was found to decrease with increasing pressure and depth, so that wollastonite is formed, e.g., at 350° at about 30 km. depth. If the existing partial $CO_2$ tension in the earth's crust is less than in the atmosphere, increasing pressure and depth will depress the reaction temperature of equation (16) more steeply. However, gravitative chemical stability exists when the

partial $CO_2$ tension increases with depth according to the law of equilibrium (2). In this case it is more difficult to compute how increasing depth affects the formation temperature of wollastonite. However, the fact that wollastonite has a greater density than quartz or calcite indicates that, at ideal gravitational stable conditions, the temperature of formation of wollastonite decreases somewhat as depth increases.

The comparatively rare occurrence of cavities filled with $CO_2$ gas in regional metamorphic marbles carrying lime-silicates indicates that $CO_2$ commonly is able to diffuse away from the place of reaction and that the processes described above actually take place in the crust under regional metamorphic conditions.

According to the calculations of V. M. Goldschmidt (1912), who regarded wollastonite as an important geological thermometer, the lack of wollastonite in regional-metamorphic marbles should indicate a temperature lower than 500° at low pressure, and below 1,000° at high pressure. However, according to the calculations above, the lack of wollastonite in most regional-metamorphic marbles proves that the temperature has been less than 500° at low pressure and perhaps less than 400° at high pressure. This indicates that regional granitization and intense metamorphism commonly take place below 400°–500°. This is in harmony with the steadily accumulating data (Ingerson, 1937) pointing toward low temperatures of granitization and related phenomena, temperatures that are lower than the solidus point of a hydrous granitic magma (Goranson, 1931).

N. L. Bowen (1940) has treated several carbonate-silicate reactions from the viewpoint that $CO_2$ is unable to diffuse away from the place of reaction. Consequently, he found the higher the rock pressure, the higher the temperature required for the formation of the silicates. However, the transformation temperatures of the reactions treated by Bowen must also be more or less independent of pressure and depth, or perhaps decrease with pressure, because $CO_2$ is actually able to diffuse through the solid rocks.

## GRAVITATIVE STABILITY OF CHEMICALLY INERT AND IMMISCIBLE CONDENSATES

Let us consider the case of two solids, *A* and *B*, as immiscible and chemically inert phases. The density of *A* is less than that of *B*. It is easy in this case to prove that the chemically stable state is established if the light *A* phase rests on the more dense *B* phase (Ramberg, 1944*a*). If *B* rests on *A* or if the two phases are mechanically mingled, then chemical forces exist which tend to diffuse *A* atoms upward and *B* atoms downward until the stable arrangement is reattained.

However, if the power of diffusion of the *A* and *B* atoms differs, then mechanical or bodily motion of the *A* or *B* crystals will also take place. If, for instance, *A* migrates upward more rapidly than *B* diffuses downward, then *B* crystals, or the whole *B* layer, sinks down through the *A* phase at the same time that *A* atoms diffuse up through the sinking *B* layer. On the other hand, if *B* diffuses more rapidly than *A*, then the downward-diffusing *B* atoms consolidate at the bottom of the system, and the expanding crystalline *B* aggregate elevates the overlying *A* crystals because of the force of crystallization of the substance *B*.

This theoretical example becomes of geological interest if we substitute for the crystalline *A* and *B* substances the light granitic and the heavier gabbroic rocks, respectively. Dense gabbroic rocks rest-

ing on light granitic rocks or magma represent a chemically unstable arrangement, so that the atoms, ions, or molecules of the granitic minerals or melt become activated and tend to diffuse up through the gabbroic rocks, which, owing to low diffusibility of the constituent elements, sink bodily.

Thus we see that the direction of the driving force of radial diffusion in the earth is in complete harmony with geological experience with regard to regional granitization. Pentti Eskola (1932), recognizing that the granitic shell was not an original feature of the earth but was formed subsequently by radial motion of substances, explained regional granitization by upward flow of light, granitic pore "magmas" through confining rocks of greater density. However, in several cases it is impossible to assume that hydrous granitic pore liquids have been present during granitization because the existing temperatures have been below the liquidus field of hydrous magmas. In other cases, also, a motion of such liquids through the very minute pore system of solid rocks is impossible. Nevertheless, field observations prove that granitic material has wandered through the solid rocks, altering them into successively more granitic types. In such cases it is important to remember that potential chemical forces are induced by mechanical instabilities and, consequently, that diffusion will account for the transportation of substances.

## SUMMARY AND CONCLUSIONS

In this paper it is shown that forces exist which tend to cause elements to diffuse in directions consistent with the present radial distribution of those elements in the earth. Consequently, one of the main objections to the theory of radial diffusion in the solid crust is annulled, viz., that the driving force of diffusion is not parallel to the direction along which matter actually has been transported. P. Niggli (1941, pp. 17–18), for example, sought these "mysterious" forces when he criticized the view of Van Bemmelen (1935), who unfortunately had only a vague impression of the forces underlying diffusion.

With reference to the velocity of diffusion through solid rocks, we refer the reader to the several geological and geochemical phenomena which definitely prove that matter is able to diffuse over shorter or longer distances throughout the crust.

The influence of gravity on the vertical diffusion and the large-scale metamorphic differentiation of our globe may be summarized as follows:

1. In a homogeneous phase (mixed crystal, magma) elements with greater fictive density (p. 451) than the density of the phase will be gradually concentrated at greater depths. Elements with smaller fictive density will be concentrated at higher levels.

2. In heterogeneous systems with different phases, elements with great fictive density are commonly discontinuously enriched downward, and less dense elements are discontinuously enriched toward the top, the several phases being arranged in concentric layers in the crust.

3. Interionic bonding forces, however, will disturb this gravitative arrangement. Elements with small fictive density but strong attraction to dense phases, for example, may be absorbed in this phase at great depth. On the other hand, elements with high fictive density but with strong attraction to less dense phases may concentrate in these light phases at the top of the crust.

## REFERENCES CITED


BACKLUND, H. G. (1946) The granitization problem: Geol. Mag., vol. 83, pp. 105–117.

BARRER, R. M. (1941) Diffusion in and through solids, Cambridge, The University Press.

BARTH, T. F. W. (1948*a*) The distribution of oxygen in the lithosphere: Jour. Geology, vol. 56, pp. 41–49.

——— (1948*b*) The Birkeland granite, a case of petroblastesis: Comm. géol. Finlande Bull., in press.

BEMMELEN, R. VAN (1935) Über die Deutung der Schwerkraftanomalien in Niederlandisch-Indien: Geol. Rundschau, vol. 26, pp. 199–226.

BOWEN, N. L. (1940) Progressive metamorphism of siliceous limestone and dolomite: Jour. Geology, vol. 48, pp. 225–274.

BROWN, HARRISON, and PATTERSON, CLAIRE (1948) The composition of meteoritic matter: III. Phase equilibria, genetic relationships, and planet structure: Jour. Geology, vol. 56, pp. 85–111.

ESKOLA, P. (1932) On the origin of granitic magmas: Min. pet. Mitt., vol. 42, pp. 455–481.

GOLDSCHMIDT, V. M. (1912) Die Gesetzen der Gesteinmetamorphose: Vidensk selsk. Oslo, Math.-naturh. Kl. no. 22.

——— (1943) Oksydasjon og reduksjon i geokjemien: Geol. fören. Stockholm Förh., vol. 65, pp. 84–85.

GORANSON, R. W. (1931) The solubility of water in granite magmas: Am. Jour. Sci., vol. 22, pp. 481–502.

INGERSON, E. (1947) Liquid inclusions in geologic thermometry: Am. Mineralogist, vol. 32, pp. 375–388.

JOST, W. (1937) Die chemische Reaktion, Vol. II, Dresden and Leipzig, T. Steinkopff.

KUHN, W., and RITTMAN, A. (1941) Über den Zustand des Erdinnern und seine Entstehung aus einem homogenen Urzustand: Geol. Rundschau, vol. 32, pp. 215–256.

MACINNES, D. A. (1939) The principles of electrochemistry, New York, Reinhold Pub. Co.

MARSH, J. S. (1935) Principles of phase diagrams, New York and London, McGraw-Hill Book Co.

NIGGLI, P. (1942) Das Problem der Granitbildung: Schweizer. min. pet. Mitt., vol. 22, pp. 1–84.

OFTEDAL, I. (1947) Remarks on the paper by H. Ramberg: Relation between external pressure and vapor tension of compounds, and some geological implications: Norsk geol. tidsskr., vol. 26, pp. 221–222.

PERRIN, R., and ROUBAULT, M. (1939) Le granite et les réactions à l'état solide: Services carte géol. l'Algérie Bull., ser. 5, no. 4.

RAMBERG, H. (1944*a*) Relation between external pressure and vapor tension of compounds, and some geological implications: Norsk vidensk.-akad. Oslo, Math.-naturh. Kl., no. 3.

——— (1944*b*) The thermodynamics of the earth's crust I: Norsk geol. tidsskr., vol. 24, pp. 98–111.

——— (1945) The thermodynamics of the earth's crust II: *ibid.*, vol. 25, pp. 307–326.

——— (1946) Kjemisk likevekt i gravitasjonsfeltet og dens betydning for jordskorpens differensiasjon: Dansk. geol. Fören. Medd., vol. 11, pp. 12–29.

——— (1947) The force of crystallization as a well definable property of crystals: Geol. foren Stockholm Förh., vol. 69, pp. 189–194.

WEGMANN, C. E. (1935) Zur Deutung der Migmatite: Geol. Rundschau, vol. 26, pp. 305–350.

# THE FORM AND STRUCTURAL FEATURES OF APLITE AND PEGMATITE DIKES AND VEINS IN THE OSI AREA OF THE NORTHERN PROVINCES OF NIGERIA AND THE CRITERIA THAT INDICATE A NONDILATIONAL MODE OF EMPLACEMENT[1]

B. C. KING
University, Glasgow, Scotland

## ABSTRACT

Numerous dikes and veins of aplite and pegmatite in the Osi area of Nigeria were studied with a view to ascertaining the mechanism of emplacement. Offsetting of the invaded rocks of a magnitude and direction appropriate to dilation is so rare as to appear to be quite fortuitous. Where offsetting is not entirely absent, it is commonly of a different magnitude and frequently in the opposite direction from that to be expected on dilation of the invaded rocks. In such instances emplacement has evidently occurred along planes of shearing formed under compressional stress.

Nondilational emplacement is also clearly indicated in the case of the irregular bodies and ramifying networks of veins. The criteria advanced are similarly of significant application in the mode of emplacement of "ptygmatic folds."

Processes of metasomatic replacement, conveniently termed "aplitization" and "pegmatization," are believed to have been those responsible for the development of the bodies described, while internal evidence for the operation of such processes is sometimes provided by the textural features of the dikes themselves.

The broader petrogenetic implications emphasize the need for a closer scrutiny of the criteria that are commonly accepted as indicative of intrusive relations.

## INTRODUCTION

Goodspeed (1940) showed that two categories could be recognized among dikes, depending on their mode of emplacement by dilation or replacement of the host rocks. The textural and structural features by which the differing modes of emplacement may be recognized were summarized (pp. 194–195).

Among the p[illegible]al field criteria discussed by Goodspeed is the evidence provided by the oblique intersection of two dikes (pp. 189–191). Emplacement by dilation of the later of the two dikes is indicated where the earlier of the dikes shows offsetting of appropriate magnitude and direction (fig. 1, *A*). Where such offsetting is absent and the disconnected portions of the earlier dike are in complete alignment, the possibility that the later dike is of dilational type is exceedingly remote and demands the postulation of a lateral displacement that precisely corresponds to the expected offsetting. The intersection of dikes without offsetting is regarded by Goodspeed as incontrovertible evidence for a replacement origin of the later dike (fig. 1, *B*); yet such intersections have been figured without comment on the mechanical difficulties involved in a dilation interpretation. An interesting illustration in this respect is figure 10 in *The Geology of Lizard and Meneage* (Flett, 1946), which shows a dike of epidiorite obliquely cutting a dike of gabbro pegmatite, which lies within a band of gabbro schist that traverses serpentine. The epidiorite also traverses a small boss of troctolite. Not one of these intersections shows any sign of offsetting. It is clear, nevertheless, that no mechanism other than dilation and injection has been envisaged for the emplacement of the several bodies; but it is only fair to observe that the

[1] Published by permission of the Director, Geological Survey of Nigeria. Manuscript received May 3, 1948.

figure is essentially diagrammatic and is primarily intended to illustrate age relationships.

## APLITE AND PEGMATITE DIKES IN THE OSI AREA OF NIGERIA

During the course of recent field studies in an unusually well-exposed area of basement complex in Nigeria, the writer examined numerous aplite and pegmatite dikes and veins. In outline, the geology of the area[2] consists of banded gneisses, granitic gneisses, and subordinate biotite- and hornblende-schists which have been involved in two granite cycles. The first was expressed as granitization *in situ* and produced a foliated granite-gneiss: during the later cycle a large body composed predominantly of coarse-grained porphyritic granite was emplaced, although in this case only part of the mass represents granitization *in situ*. A number of sheeted bodies, now consisting of metagabbro, invaded the area between the two granite cycles.

Apart from numerous semiconformable, medium-grained leucocratic veins, many of which undoubtedly antedate even the granite-gneiss, dikes and veins of both aplite and pegmatite were formed in great abundance during both the granite cycles. In summary these may be grouped as follows:

7. latest pegmatites
6. latest aplites
5. first postporphyritic granite pegmatites
4. first postporphyritic granite aplites
3. preporphyritic granite pegmatites
2. postgabbro aplites and pegmatites
1. postgranite-gneiss aplites and pegmatites

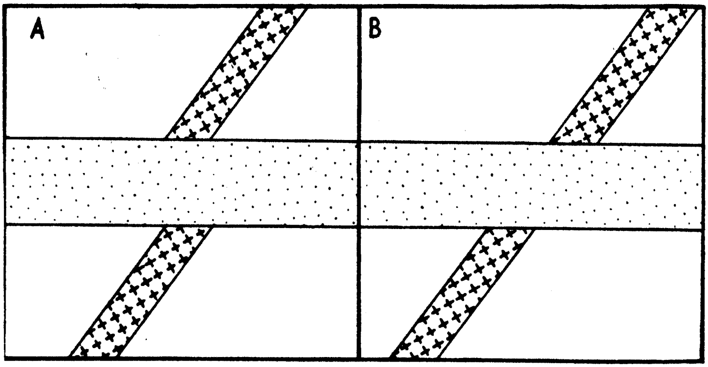



FIG. 1.—Diagram illustrating the oblique intersection of dikes. In *A*, offsetting of the earlier dike takes place, indicating a dilational mode of emplacement of the later dike. In *B*, no offsetting occurs, so that emplacement of the later dike must have taken place without dilation of the dike walls.

Figures 2, 3, 4, and 5 illustrate typical field characters of these groups of aplites and pegmatites. The absence of offsetting at intersections appropriate to emplacement by dilation is to be observed in a number of instances (figs. 3 and 4, *A*, *B*, and *C*). In addition, especial attention is directed to the following phenomena: (*a*) the commonly highly irregular form of the bodies, with "pinches" and "swells" that produce no effect on the structure of the adjacent host rock (fig. 2, *C*, etc.); (*b*) the enclosure of undisturbed patches of the invaded rock, in many places so extensive that they occupy the greater part of the width of the

[2] A detailed account of the geology is in preparation for publication by the Geological Survey of Nigeria.

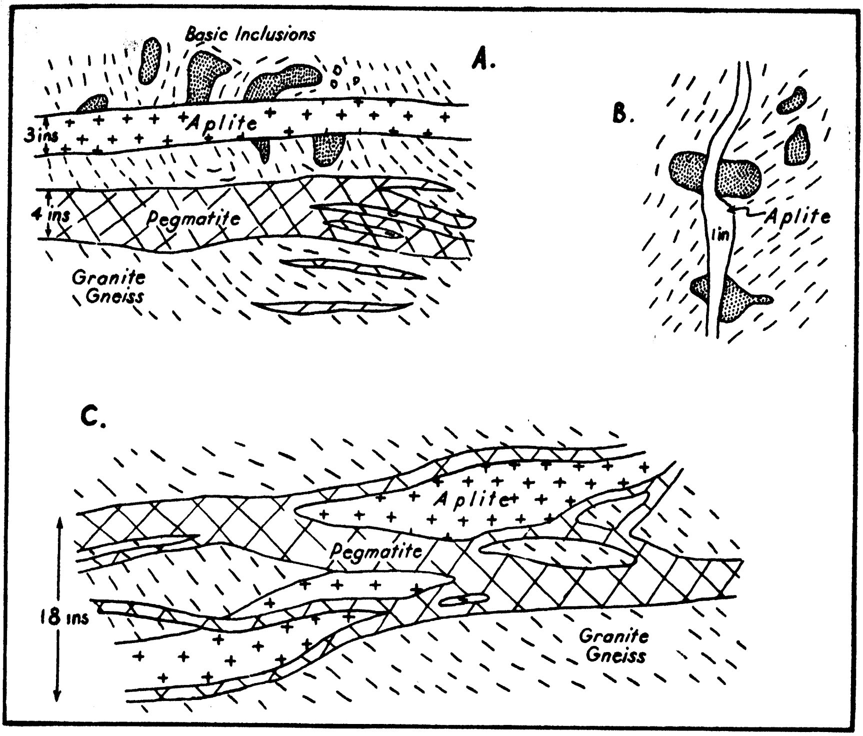



FIG. 2.—Aplites and pegmatites invading granite-gneiss, from Osi area of Nigeria. Note irregular form of pegmatite dikes and the absence of corresponding parts of basic inclusions.

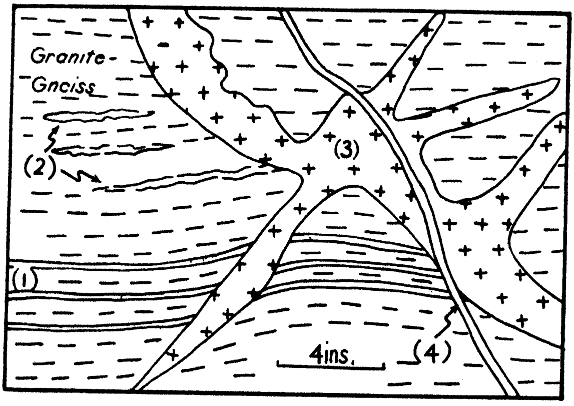



FIG. 3.—Aplites and pegmatites invading granite-gneiss, from Osi area of Nigeria. *1* represents bands in original banded gneiss, which have been partly "rejuvenated" during granitization. *2* shows streaks in granite-gneiss, due to extreme felspathization, associated with granitization. *3* represents aplite related to latest phase of granite-gneiss cycle. *4* shows pegmatite veins related to later porphyritic granite cycle.

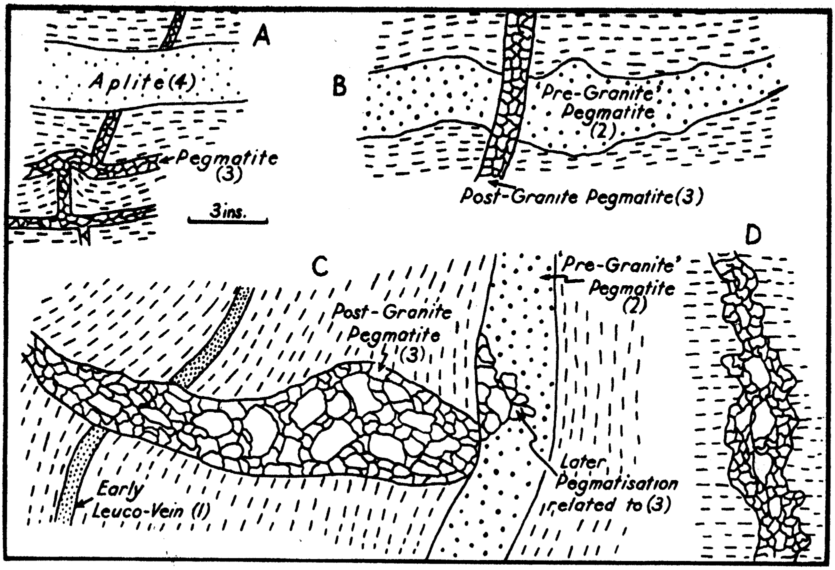



FIG. 4.—Relations between aplites and pegmatites in granite-gneiss, from Osi area of Nigeria. The pegmatite in *A* shows displacement of greater magnitude and in the wrong direction from that required by dilation of the walls of the aplite. *D* shows the character of a pegmatite transverse to the foliation in the granite-gneiss and its evident replacement habit. All dikes belong to the porphyritic granite cycle.

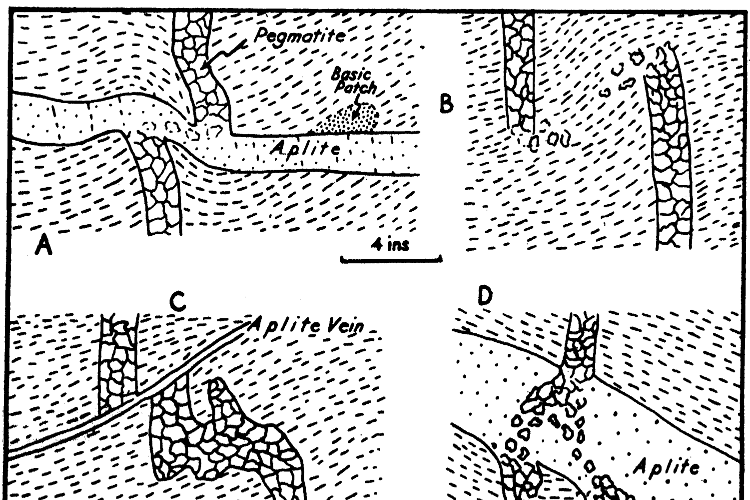



FIG. 5.—Features of aplites and pegmatites belonging to the porphyritic granite cycle and invading granite-gneiss. In all diagrams there is evidence of lateral displacement affecting the pegmatite dikes or the fissures to which they were related. In *A* and *B* there is evidence of compressional stress leading to plastic distortion of the granite-gneiss foliation. Offsetting caused by displacement is partly bridged by pegmatization of the intervening host-rock.

dike (fig. 2, *A* and *C*); (*c*) the transition of regular dikes into a series of tongues (fig. 2, *A*) or into a body with indefinite margins; and (*d*) the truncation of inclusions and basic patches in the host rock and the absence or lack of correspondence of related portions within the dike or on the opposite wall (figs. 2, *A* and *B*, and 5, *A*).

The inadequacy of dilation even as a partial mechanism for the production of the observed features cannot be overemphasized: its application in the case of the "star-shaped" body depicted in figure 3 would be particularly improbable; and, moreover, it is to be borne in mind that the addition of the third dimension serves further to intensify the difficulties that are already apparent in the application of a dilation mechanism to the present two-dimensional illustrations.

A number of the diagrams indicate the operation of a tectonic process that is incompatible with the physical environment within which dilation is likely to be associated. This is the development of shearing effects in the rocks in response to regional compressional stress. It is not proposed here to detail the evidence for the operation of such stress on a regional scale, since it is intended that this aspect will form the subject of a later contribution. In the present account attention will be confined to the relation between stress movements and the mode of emplacement of the dikes.

Observation of numerous examples of displacements of earlier by later dikes showed that the number of cases in which the displacement is of such magnitude and direction as to accord with the offsetting appropriate to dilation are so rare as to suggest that they are quite fortuitous. The actual displacements found, by comparison with those required by dilation, are generally too great, usually in the opposite direction; are unrelated to varying dike width; and commonly vary not only in magnitude but even in some cases in direction, along different parts of the same dike. Figure 4, *A*, shows that shearing movement, which apparently determined the trend of the aplite dike, has been responsible for the displacement of the earlier pegmatite. In figure 5, *C*, the relative positions of the offset portions of the pegmatite dike show that displacement occurred in response to compressional stress. The shear plane itself became the locus of an aplite vein. Precisely similar relations may be observed in displacements of structural features in the host rock, where these are nonconformably traversed by dikes.

Although it is apparent that in many instances the host rocks have undergone brittle displacements as a result of shearing stress, with the formation of fractures that have commonly been the loci along which aplite or pegmatite dikes and veins were subsequently emplaced, it is interesting to observe that in other cases the rocks have undergone plastic deformation. Indications of the latter are seen where the relative lateral displacements on either side of a dike show rapid variations both in magnitude and in direction. No simple mechanical explanation of such features is possible, for the rocks have behaved as if they were capable of localized compression and elongation.

Related features are seen in figures 4, *A*, and 5, *A*, where, near the intersection of two dikes, the invaded granite-gneiss has been affected by what may be described as "plastic rotation" along an axis perpendicular to the plane of the diagrams. In figure 5, *A*, especially it is apparent that the gneiss underwent rotation that is relatively greater than that

shown by the aplite dike, a feature which suggests that deformation of the gneiss commenced before emplacement of the dike. The pegmatite (or its predetermining fracture) has experienced a corresponding displacement on either side of its intersection with the aplite. Similar plastic deformation induced by compressional stress operating at right angles to the plane of foliation is also demonstrable in figure 5, *B*. Dislocation has been distributed through a zone and is not confined to a definite shear plane. As in *A* and *C* of the same figure, compression is indicated by the "overlapping" of the displaced ends of the dike.

Figures 6 and 7[3] are of interest in that they depict "anomalous" features of otherwise regular and persistent large aplite dikes. Figure 6 shows part of a large dike, which, were it not for the fact that it is traversed in one place by an offshoot of the host rock, would be regarded as having been emplaced within the latter. Actually, the formation of the dike preceded that of the surrounding granite. The field evidence demonstrates that the dike was emplaced within the coarse porphyritic granite which forms the greater part of the granite mass. Later this granite was locally replaced by a more even-grained leucocratic granite. Relics of the earlier granite are now represented only by inclusions within the aplite dike, which at one point (not shown in diagram) consists merely of narrow septae and marginal selvages enclosing undisturbed blocks of the earlier granite. The unrotated character of the inclusions is apparent from the conformance of the characteristic "flow-layering" of the feldspar phenocrysts.

The features illustrated in figure 7 would place quite extraordinary demands on a hypothesis of dilation with concomitant injection of aplite. Both apophyses are continuous with the same larger dike, yet one obliquely cross-cuts the other without effecting displacement.

[3] The original photographs from which figs. 6 and 7 were prepared will be reproduced in the publication by the Geological Survey of Nigeria.

## ANALOGOUS FEATURES FROM OTHER AREAS

In an account of the Cnoc nan Cuilean area of the Ben Loyal complex of Sutherland (King, 1943), the writer employed two diagrams (figs. 1 and 2) illustrating dikes and veins of aplite and pegmatite which traverse the variable syenites of the marginal zone of the body. These diagrams have been redrawn as figures 8 and 9 in this paper. At the time of their original publication, the writer observed nothing in the mutual relations of the dikes which might reflect on their mode of emplacement. It is now apparent that they are beautiful examples of the absence of offsetting of earlier by later obliquely intersecting dikes. Furthermore, the absence of corresponding portions of truncated "basic patches" from the opposite sides of small dikes are even more clearly shown than are similar features from the Osi area. The significance of parallel sides to undulating veins will receive attention later (p. 469).

Comparable features in aplite dikes were also recorded by the writer during the course of mapping in the Kawungera area of central Uganda during 1945, although it is only in the light of the examples from the Osi area that the significance of the Uganda occurrences has been appreciated. In the Kawungera area a mass of coarse-grained porphyritic granite invades phyllites, sandy phyllites, and quartzitic sandstones of the Karagwe-Ankolean system but is older than the Singo series, which here consist of a thin basal sandstone followed by

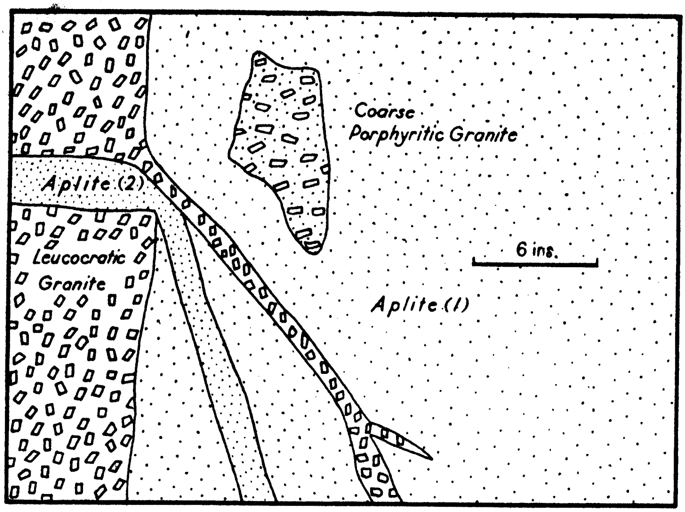



FIG. 6.—Part of a large aplite dike within coarse leucocratic granite: a late phase of the porphyritic granite. For interpretation see text, p. 464. From Osi area of Nigeria. (Drawn from a photograph.)

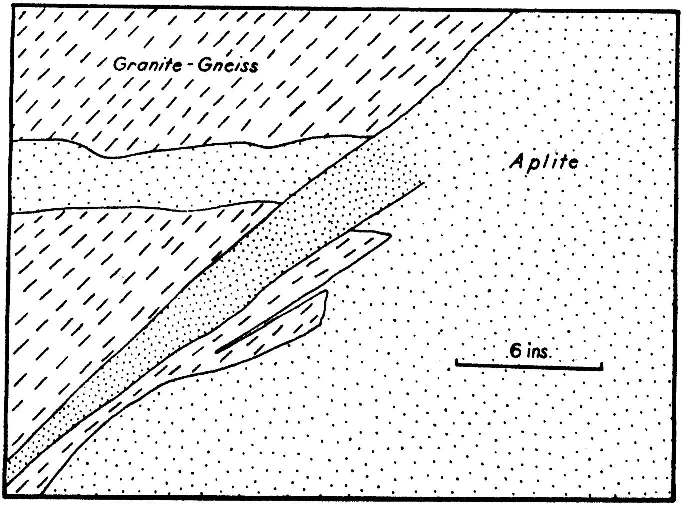



FIG. 7.—Margin of large aplite dike cutting granite-gneiss, showing mutual relation between two apophyses. Although both pass continuously into the main dike, one is cross-cutting toward the other. From the Osi area of Nigeria. (Drawn from a photograph.)

silicified "siltstones" (King, 1947, pp. 38–40). Aplite dikes and veins and a few pegmatites traverse the granite; the accompanying diagrams (figs. 10 and 11) illustrate some of the features of these aplites. In figure 10, *A*, the partly irregular form of the aplite dike and associated veins, the main trend of which is apparently determined by parallel fractures, may be compared with that of the composite aplite-pegmatite dike (fig. 2, *C*) from the Osi area. Figure 10, *C*, provides an example of offsetting in the opposite direction from that demanded by dilation of the dike walls. In figure 11 it is clear that, if the smaller body had been emplaced by dilation-injection, the width of the larger dike would have been the same on both sides of the later body.

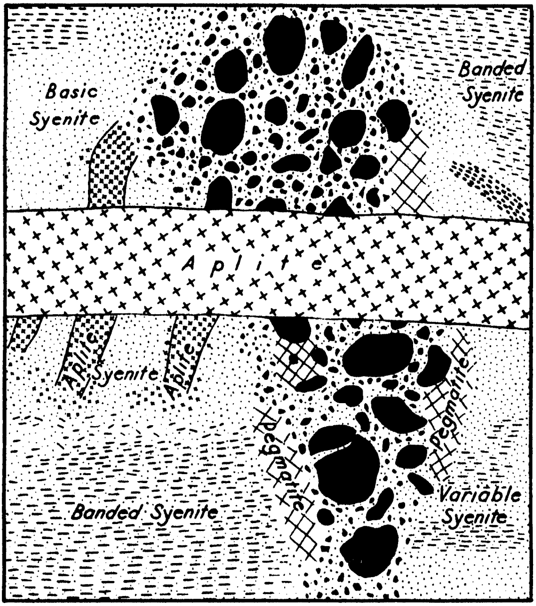



FIG. 8.—Aplites in "variable" syenites from Cnoc nan Cuilean area of Sutherland. The large aplite (which is about 4 inches across) truncates "basic patches" and earlier aplites; but corresponding portions are not represented on opposite sides of the dike. Areas of diffuse aplitization and pegmatization may also be observed. Redrawn from Fig. 1 of Vol. 98, Quart. Jour. Geol. Soc. for 1943.

### RAMIFYING NETWORKS OF DIKES OR VEINS

A notable feature of Archean terrains is the presence of irregular networks of anastomizing dikes and veins of aplite, pegmatite, or quartz. The sheets may conform partly or wholly to certain trends or may be entirely irregular, depending, apparently, on the presence or absence of directional structures within the host rock along which planes of weakness or fracture may develop preferentially. Similar bodies are also found, though generally more sporadically, in relation to granite plutons of later ages.

Little attention appears to have been paid to the mechanism by which the host rocks have accommodated themselves to the emplacement of such complex bodies,

which commonly belong not merely to one but to a number of epochs. Figures 12 and 13 illustrate typical assemblages of such veins from the Osi area. Figure 9 presents similar features from another region. Attention is directed to the following characters: (*a*) the persistence of many of the veins and their comparatively constant width, despite numerous intersections and the numbers of different trends that are represented; (*b*) the many examples of mutual intersections without offsetting of either vein; (*c*) the continuity of structure in the various isolated blocks of host rock; and (*d*) the parallelism of the walls of those veins that have undulating forms as well as of those that are straight.

In order to demonstrate the effect on host rocks of uniform tensional stress necessary to accommodate a typical network of veins, figures 14 and 15 were drawn to illustrate the relative positions of blocks of host rock both before and after uniform dilation has operated along a given set of fractures. The following features of the diagrams are to be noted: (*a*) the longer and straighter fractures do not dilate in such a fashion that they could be occupied by regular veins of constant width; (*b*) intersections almost invariably show offsetting of a magnitude that depends on the relative size and shape of the adjacent blocks; (*c*) formerly continuous structures in the host rock become displaced; and (*d*) precise parallelism of the walls of undulating fractures is never maintained after dilation—the disparity increases with increasing curvature of the vein.

It is readily apparent that the resultant picture in no respect accords with the features actually observed in the field. Comparable patterns may perhaps be confined to examples of brecciation that has occurred at relatively shallow depth. In particular it should be noted that a dilation mechanism is almost incapable of producing tongue-like apophyses which do not completely traverse isolated blocks, such as are almost invariably found in actual examples of vein networks (fig. 12).

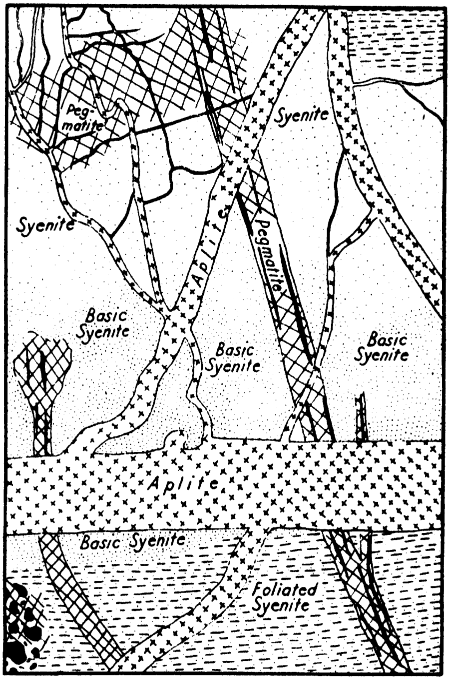



FIG. 9.—Network of aplite and pegmatite dikes and veins in "variable" syenites from the Cnoc nan Cuilean area of Sutherland. Numerous intersections occur without offsetting, while the general pattern of the ramifying veins is not that found with dilation (see also p. 465 of text). Redrawn from Fig. 2 of Vol. 98, Quart. Jour. Geol. Soc. for 1943.

The general principle of the above demonstration is not affected if we consider the tensional stress to predominate in one direction; in such cases a square of the original pattern will be rectangular or rhomb-shaped after dilation, whereas, even if the host rocks are regarded as capable of nonuniform dilation, compen-

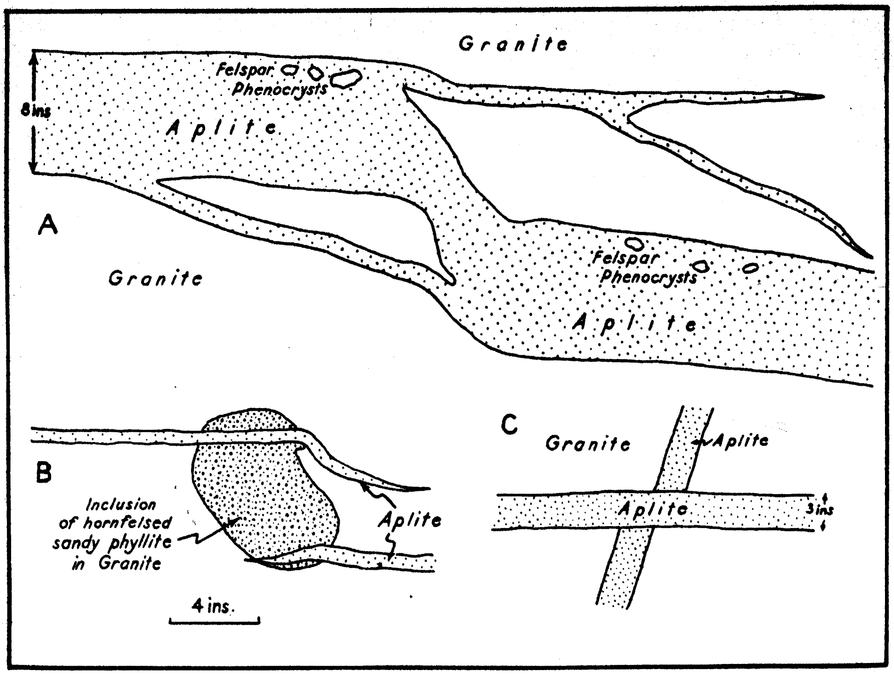



FIG. 10.—Aplites from the Kawungera granite area of central Uganda. *A* shows the development of a dike along two parallel fractures. *B* illustrates a similar feature on a smaller scale, in which aplite veins cut an inclusion in granite without displacement of the walls. *C* shows offsetting of an earlier dike by a later of larger magnitude and in the opposite direction from that expected to result from dilation.

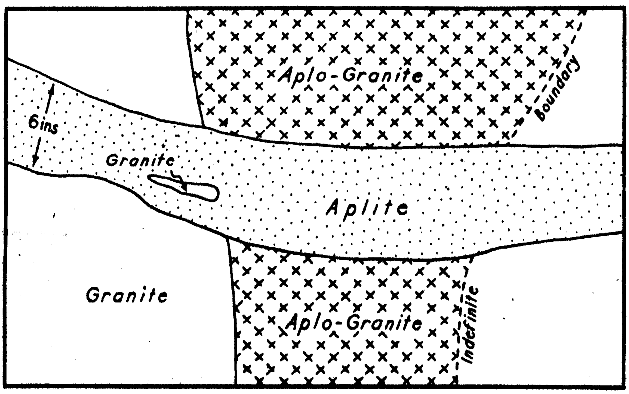



FIG. 11.—Aplite and aplo-granite, cutting granite in the Kawungera area of central Uganda. The aplite dike has evidently developed without dilation of the host-rock.

sation of the numerous local irregularities must occur over larger units. Again, it may be noted that, whether the dikes and veins belong to one or to more than one period of formation, the general argument remains the same.

partings only at the crests of the plications, whereas those portions of the fracture that are to be occupied by the subparallel limbs of the vein folds will not open at all.

Read (1928) and, more recently, Bud-

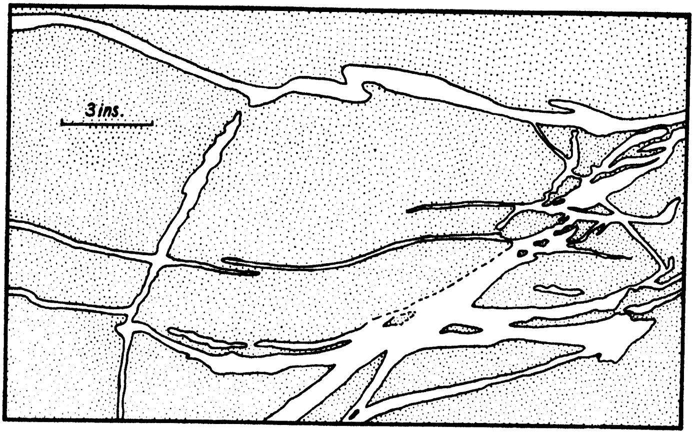



FIG. 12

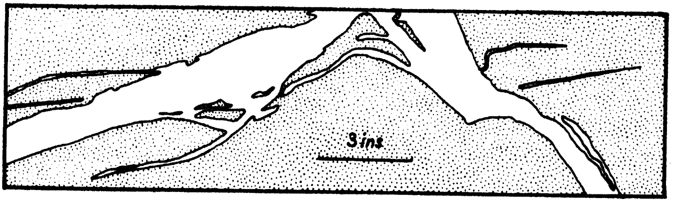



FIG. 13

FIGS. 12 and 13.—Examples of ramifying networks of veins of aplite composition and texture which invade partly granitized metagabbro in the Osi area of Nigeria. The margins of the veins are often indistinct, while diffuse patches of similar composition also occur in the host-rock. (Drawn from photographs.)

### PTYGMATIC VEINS

It is readily apparent by experiment that a vein will not have parallel sides if it has been emplaced by simple injection between the dilated walls of an undulating fracture. In the case of ptygmatic veins the difficulty is especially acute, since it is easily verifiable that dilation in a general direction at right angles to the main trend of such a vein will cause

dington (1939, pp. 164–165) state that ptygmatic veins have their original form and were never planes. Read (p. 75) in a number of sketches amply demonstrates this contention—a conclusion that is the more incontestable because it is evident that the foliation in the invaded Moine series largely agrees with the original sedimentary bedding. Nevertheless, Read maintains "the clearly intrusive

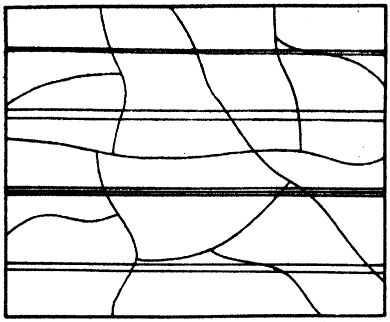


FIG. 14

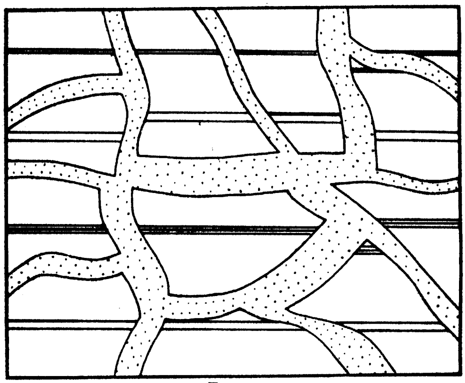


FIG. 15

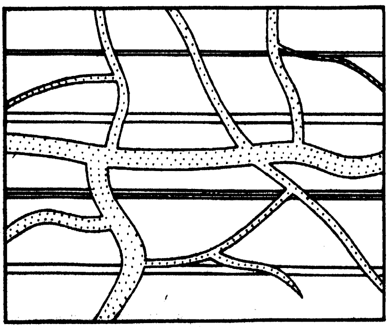


FIG. 16

FIGS. 14–16.—Fig. 15 illustrates the effect which may be expected from uniform dilation along the fractures indicated in Fig. 14. Noteworthy features are the displacements of original structures and of intersections; the lack of uniform width to channels following the longer fractures; and the absence of parallel sides to channels along curving fractures. In Fig. 16 replacement is imagined to have operated along the same set of fractures. The correspondence between the features so produced and those found in field occurrences of networks of dikes and veins is to be noted.

igneous nature of the veins, and their definite association with great igneous bodies of unquestionable origin." The tortuous character of the veins is ascribed to the "absence of easily opened plane channels for the injected material."

The evidence presented in this paper, in its application to ptygmatic veins, leads clearly to the conclusion that, if their mode of emplacement was by dilation-injection, their present form must have been attained by deformation of original flat sheets; whereas, if it can be proved that the latter was not the case but that their present form is an original character, then the veins have not been formed by dilation with concomitant injection.

The writer, although not contesting the evidence that ptygmatic veins, in the examples given by Read, owe their form to an original character, has noted examples in which similar bodies have evidently been produced by deformation of sheets of simpler form. In figure 17 the close conformity of many of the tortuous veins with relic structures indicates that they were originally semiconformable veins in the banded gneisses, but the axes of the minor plications now accord with the impress of a later deformation. "Rejuvenation" of the earlier veins and further accretion of leucocratic material have taken place during the granite-gneiss cycle, to which the larger irregular patches are to be solely ascribed. The two more regular aplite dikes seen in the diagram are related to a closing phase of the later cycle, and their general conformability to the granite-gneiss foliation is apparent.

## THE MODE OF EMPLACEMENT OF NON-DILATIONAL DIKES AND VEINS

Because it is demonstrable that the form displayed by the dikes and veins which have been described above could

not have resulted from the dilation of fracture walls accompanying the injection of the dike-filling material, it is apparent that the volumes now occupied by the dikes and veins were formerly occupied by host rock. The latter must therefore have undergone replacement or displacement. Whereas, in the case of larger bodies, it is not necessarily easy to preclude the possibility that the invaded composition (such as quartz-feldspar aplite within biotite- or hornblende-schist). The usual absence of signs of "contamination" is inconceivable if bodily removal of the invaded rock is postulated. The transformations must, therefore, have been effected in terms of ultimate crystal units by processes of metasomatic replacement, which, in order more specifically to imply the uniformity of the


FIG. 17.—Complex system of veins in partly granitized banded gneisses. The veins are essentially of aplite but are somewhat pegmatitic in places and are largely based on semiconformable leucocratic bands in the earlier gneiss, which were crumpled and "rejuvenated" during the granite-gneiss period of deformation. The original conformable structure is apparent where the relic foliation of the banded gneisses is largely retained (see especially lower center). Later granitization is tending to produce a new foliation parallel with the axes of the minor folds: this direction is also followed by the two dikelike bodies of aplite which belong to the later phases of the granite-gneiss cycle. The development of the larger aplite areas is to be ascribed to the later cycle, while "rejuvenation" has caused the dikes to appear as continuous with the contorted veins. From the Osi area of Nigeria. (Diagram drawn from a photograph.)

rocks have disappeared by a process of disintegration and mechanical removal, such as "stoping" (Billings, 1925), for small and irregular dikes or ramifying networks of veins this is virtually impossible. The vein networks are often essentially "mechanically closed" systems and may be completely so in the two dimensions in which they can normally be studied. In some cases, too, the veins and host rock are of strongly contrasting resultant products, may be designated as "aplitization" or "pegmatization."

It is clear that the aplitizing or pegmatizing agencies operated initially along fractures or other planes of weakness within the host rocks. In strongly banded or foliated rocks semiconformable veins are commonly found, but the larger dikes tend to be disposed along shear planes in areas where these are present. More irregular dikes and veins form if the host

rock is deficient in preferred directional structures or if, in banded or foliated rocks, the dikes and veins have formed in directions that do not accord with natural planes of easiest fracture. Highly irregular margins, evidently ascribable to the effects of uneven replacement, are commonly characteristic of such bodies, of which examples may be seen in figure 4, *D* (pegmatite formed across the foliation of granite-gneiss), and certain of the veins in figure 12 (weak development of preferential directions in the host rock).

processes must have produced such contacts have been described (Webb, 1946). As a further example may be cited a small dike of pegmatite which cuts porphyritic granite in the Osi area. The dike contains phenocrysts of microcline which are indistinguishable from, and possess the same orientation as, those in the adjacent granite. At the contact of the dike, certain of the phenocrysts are shared by both granite and pegmatite, despite the fact that in other respects the junction is perfectly sharp (fig. 18). The conclusion

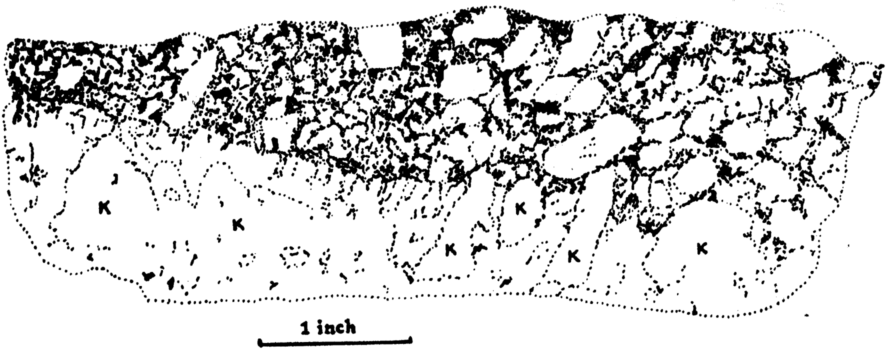



FIG. 18.—Junction between quartz-felspar pegmatite and marginal porphyritic granite from the Osi area of Nigeria (*K* 8). The microcline phenocrysts of the granite are also found within the pegmatite: some are actually shared by both rock types. In the pegmatite, further accretion of microcline has occurred, with the production of larger, more rounded plates. The principal areas of microcline in the pegmatite are indicated (*K*), but usually enclose small blebs of quartz and oligoclase.

In figure 16 a network of veins has been drawn on the assumption that they were formed by agencies of replacement which operated uniformly from the same set of fractures that were employed in figures 14 and 15. It is readily apparent that the resultant pattern and individual features of the veins accord with those that were summarized on page 466 as characteristic of networks of veins of the type under discussion.

The presence of "sharp" contacts has often acted as a deterrent toward explanations that depend on replacement, although examples in which replacement

that the pegmatite has formed essentially by a modification of the matrix of the granite in this case appears inescapable.

Those who concede that replacement may be partly responsible for the formation of dikes and veins such as those described and who retain dilation-injection as the primary mechanism in dike formation must account for the formation of a uniform rock by two entirely different processes. Moreover, they must recognize that replacement, which predominates in the marginal portions of the bodies, is responsible for the sharp contacts.

Under the microscope, even the sharp contacts that in the field are characteristic of the more regular aplite and pegmatite dikes and veins are no longer so distinctive. Instead are found the "crystal continuities" that Stillwell (1918, p. 194) regarded as indicative of "metamorphic diffusion." Individual crystals are shared at all points along the contact by both dike and host rock. In spite of Stillwell's evident confidence that "igneous" phenomena are ascribable to an entirely different mechanism, it is significant that he believed that such metamorphic diffusion took place in the solid state (1918, pp. 12, 204–208). Reynolds (1947, p. 221) has recently suggested the possibility that "sharp" contacts represent diffusion limits, and she draws attention to analogous features in experimental results on solid diffusion.

In all the areas from which examples have been cited, besides discrete dikes and veins with "sharp" contacts, there also exist less regular and more indefinite patches of aplitization and pegmatization (figs. 9 and 17). Even along a single body, variations from sharp to indistinct boundaries are not uncommon. The retention of relic structures of the host rock, apparent in the hand specimen, is a common feature of the irregular patches of aplite and pegmatite, and, under the microscope, even the regular bodies often betray an origin by replacement of the pre-existing rock. This is especially evident in certain pegmatites, wherein the matrix shows close textural and mineralogical affinities with the invaded gneisses or granite. A detailed statement of the petrographic evidence for this contention will be presented in the full account of the Osi area.

In the development of the dikes and veins under discussion, introduction of material on a considerable scale has not normally been involved. In general, the aplites and pegmatites are richer in silica, alkalies, and alumina and poorer in the basic constituents than are the gneisses or granites that form the host rock. The destination of the displaced basic constituents is not always a matter of conjecture because, most especially in the case of the irregular masses and the semiconformable veins, basification of the host rock in the vicinity of the contacts is commonly observed (fig. 19). Precisely similar features were recorded by Read, who describes "biotite selvages" to the veins of the "injection-gneisses" of Sutherland (1931, fig. 9 and compare also pl. II, *A*). A comprehensive study of the geochemistry of analogous, but larger-scale, phenomena in relation to granitization has recently been made by Reynolds (1946).

### FURTHER PETROGENETIC CONSIDERATIONS AND CONCLUSIONS

It has been demonstrated in the foregoing account that sharp contacts and transgressive relations with the host rock are by no means adequate criteria for the assumption that dikes and veins have been emplaced by a dilation-injection mechanism. In marked contrast to minor bodies that have been emplaced at higher crustal levels, aplites and pegmatites show nothing to correspond with the obvious chilled margins that characterize, for example, basic dikes. In the absence of such incontrovertible evidence of a former magmatic nature, it is necessary, therefore, to examine closely the relations between dike form and structure of the country rock before rejecting a replacement hypothesis in favor of one of dilation with concomitant injection.

The criteria on which such discrimination is possible have significant applications besides those which relate to the

mode of emplacement of dikes and veins. As an example may be cited the illustration (fig. 99, p. 271) in *Igneous Rocks and the Depths of the Earth* (Daly, 1933), which is entitled "Arrested Stoping at the Roof of the Lausitz Granite Batholith." The mechanism implied by this process is clearly indicated by Daly in the text: the blocks of country rock are regarded as having been riven apart by numerous granite apophyses and ultimately, when completely surrounded by granite magma, as becoming capable of independent relative movement.

Although it is appreciated that the details of such a diagram should not be subjected to too precise a scrutiny, it is evident that it shows none of those features which, as summarized on page 467, will inevitably be found where dilation has occurred. In particular, it may be noted that the corresponding corners of adjacent blocks, which also represent the terminations of apophyses, are in precisely the positions that they would occupy were the apophyses not there at all. The small horizontal vein near the right of the figure follows a direct course which neither affects, nor is affected by, its intersection with two other veins.

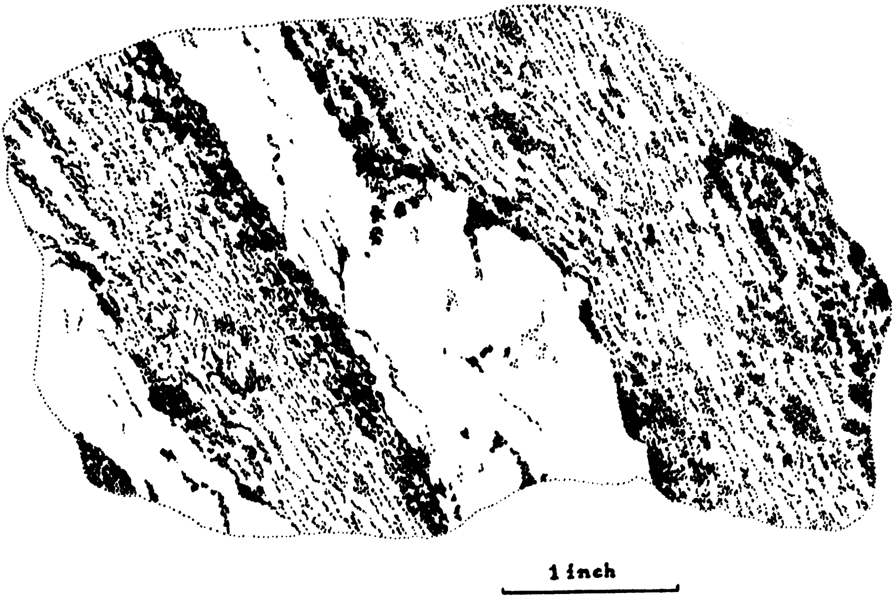



FIG. 19.—Leucocratic veins in partly granitized banded gneiss (granite-gneiss cycle) from the Osi area of Nigeria (*K* 134). Marginal basification (hornblende and biotite) adjacent to the leucocratic veins is well developed, while there is a concomitant partial disappearance of the well-marked foliation. The darker gneiss at the extreme right of the illustration indicates proximity to another vein (not shown on this side of the specimen).

The relevance of this example is the more apparent if one recalls that it was the discovery of veins of granite continuous with a main granite mass penetrating the mica schists and limestones of Glen Tilt that Hutton found especially convincing evidence for the existence of granite magma (Read, 1943, pp. 67–68).

Now that the presence of sharp contacts can no longer be regarded as conclusive evidence in this direction, the examination of the form of such apophyses and their relation with the structures of the invaded rocks has become of critical importance, and, although the absence of indications of dilation does not preclude the possibility of magmatic injection, its likelihood in such instances becomes exceedingly remote.

ACKNOWLEDGMENTS.—I wish to express my thanks to Professor Arthur Holmes, Dr. Doris Reynolds, and Professor Neville George for reading the manuscript and for their many helpful comments.

## REFERENCES CITED

BILLINGS, M. P. (1925) On the mechanics of dike intrusion: Jour. Geology, vol. 33, pp. 140–150.

BUDDINGTON, A. F. (1939) Adirondack igneous rocks and their metamorphism: Geol. Soc. America Mem. 7.

DALY, R. A. (1933) Igneous rocks and the depths of the earth, New York, McGraw-Hill Book Co., Inc.

FLETT, J. S. (1946) Geology of the Lizard and Meneage: Geol. Survey of Great Britain Mem.

GOODSPEED, G. E. (1940) Dilation and replacement dikes: Jour. Geology, vol. 48, pp. 175–195.

KING, B. C. (1943) The geology of the Cnoc nan Cuilean area of the Ben Loyal igneous complex: Quart. Jour. Geol. Soc. London, vol. 98, pp. 147–185.

——— (1947) The textural features of the granites and invaded rocks of the Singo batholith of Uganda and their petrogenetic significance: Quart. Jour. Geol. Soc. London, vol. 103, pp. 37–64.

——— and DE SWARDT, A. M. J. The geology of the Osi area of the northern provinces of Nigeria: Geol. Survey of Nigeria Bull. (In press.)

READ, H. H. (1928) A note on "ptygmatic folding" in the Sutherland granite complex: Great Britain Geol. Survey, Summary of Progress for 1927, pp. 72–77.

——— (1931) The geology of central Sutherland: Great Britain Geol. Survey Mem.

——— (1943) Meditations on granite: pt. I, Geol. Assoc. Proc., vol. 54, pp. 64–85.

REYNOLDS, D. L. (1946) The sequence of geochemical changes leading to granitisation: Quart. Jour. Geol. Soc. London, vol. 102, pp. 389–446.

——— (1947) The granite controversy: Geol. Mag., vol. 84, pp. 209–233.

STILLWELL, F. L. (1918) The metamorphic rocks of Adelie Land, sect. 1, Australasian Antarctic Expedition, 1911–1914, Adelaide.

WEBB, J. S. (1946) A replacement "pegmatite" vein in the Carn Brea granite: Geol. Mag., vol. 83, pp. 177–185.

# DEFORMATION OF QUARTZ CONGLOMERATES IN CENTRAL NORWAY[1]

CHRISTOFFER OFTEDAHL
University of Oslo

ABSTRACT

All three axes of deformed ellipsoidal pebbles have been measured in quartz conglomerates in the Sparagmite rocks (arkosic sandstones) of central Norway. More than one hundred measurements have been treated statistically. Shearing motion is considered as the main element of the deformation, and various problems concerning the deformation are discussed.

## INTRODUCTION

In orogenic regions the movement of peripheral nappes over long distances has been well known to geologists for a long time. For geological reasons rocks are supposed to have been thrust over large distances, perhaps many hundred kilometers. Such thrust movements are accompanied by deformation of the rock itself. From a geological and microscopic examination it is easy to get a rough qualitative estimate of the degree of deformation. But it is a very difficult task to get a quantitative determination of rock deformation. The only paper that I have found chiefly devoted to this problem is that recently published by Ernst Cloos (1947). By measuring the deformation of the oöids in oölitic limestones, Cloos obtained quantitative expressions for the rock deformation in the South Mountain fold of southern Pennsylvania and Maryland.

In the present study I have tried to give a quantitative expression of the deformation of sandstones by measuring the three axes of deformed pebbles in quartz conglomerates. The results are not analogous to those of Cloos because the deformation is supposed to be of a different type than that investigated by Cloos.

[1] Manuscript received November 17, 1947.

## GEOLOGY OF THE AREA

The quartz conglomerates investigated are found in the Sparagmite formation in central Norway (fig. 1). "Sparagmite" is a Scandinavian designation of certain coarse feldspathic sandstones, with a feldspar content of 20–30 per cent, rarely 10–40 per cent. In the Moelv division of the formation there are some rare quartz conglomerates. The pebbles consist of pegmatitic quartz and fine-grained quartzites. The undeformed pebbles had diameters between 3 and 10 cm. The thickness of the conglomeratic layers amounts to a couple of meters.

The Sparagmites are of Eo-Cambrian age. They were deposited on pre-Cambrian gneisses and were overlain by Cambro-Silurian strata, the thickness of which probably did not exceed 2,000 meters. During the Caledonian orogeny the Sparagmites were thrust toward the southeast. The thrust width is estimated by various authors to be between 40 or 50 and 300 km.

During the deformation the Sparagmites were slightly metamorphosed, as shown by the formation of sericite, alteration of feldspar, and the formation of a marked plane of schistosity. The schistosity is parallel to the conglomeratic layers and to minor traces of primary

stratification, such as the border between a layer rich in feldspar and a layer free from feldspar or thin dark bands of layers rich in clay. Few exceptions exist. It is believed, therefore, that no general folding occurred in these rocks.

The displacement of the rocks must have taken place in the following way: northern part of the Sparagmite area, the originally spherical pebbles were deformed to triaxial ellipsoids. Here the planes of schistosity display gentle dips (< 30°), but the strike is likely to vary from one mountain to another. The long and intermediate pebble axes lie within the schistosity planes, and the short

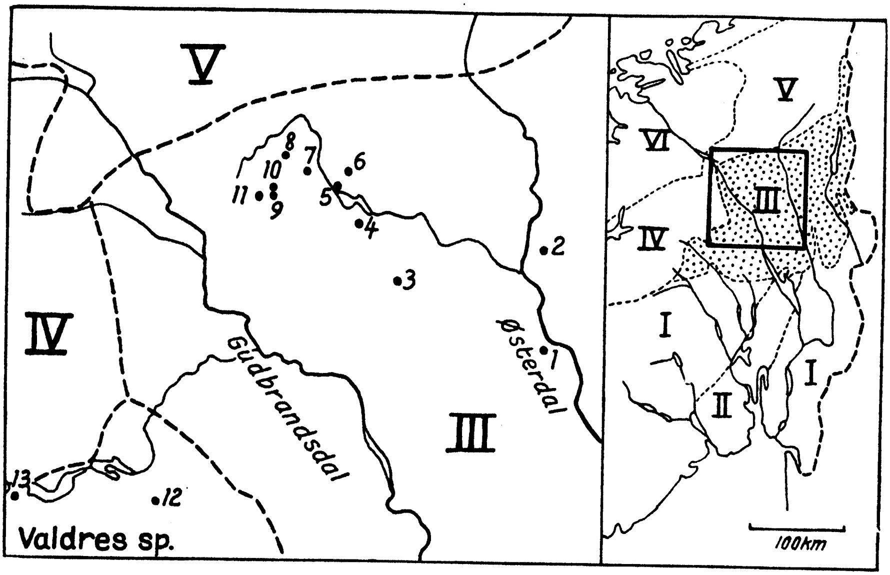



FIG. 1.—*To the right:* Key map. The formations are: *I*, pre-Cambrian gneisses. *II*, sediments (Cambro-Silurian) and igneous rocks (Permian) of the Oslo region. *III–VI*, rocks of the Caledonian orogeny. *III* (*dotted*), sparagmites. *IV*, thrust igneous rocks of Jotunheimen, etc. *V*, folded Cambro-Silurian of the Trondheim region. *VI*, highly metamorphic gneisses of the root zone.

*To the left:* Map showing localities of conglomerates, carrying pebbles which have been measured. (This area corresponds to the one inside the black square on the small-scale map.) *1*, Gråvola, 15 km. N.NW. of Koppang, Østerdal; *2*, Kivfjell, 19 km. N.NE. of Atna, Østerdal; *3*, Storfjellet, 19 km, W.SW. of Atna, Østerdal; *4*, Gråhøgda, 10 km. S. of Atnesjø; *5*, at Hyrsa, 2 km. NW. of Atnesjø; *6*, Setervola, 3 km. NE. of Atnesjø; *7*, Veslekollhø, Rondane area; *8*, top of Högronden, Rondane area; *9*, col between Rondholet and Storbotn, Rondane area; *10*, near top of Rondeslottet, Rondane area; *11*, Rondhalsen, Rondane area; *12*, northern border of N. Etnedal quadrangle; *13*, Eastern end of Lake Bygdin. "Valdres sp." is the so-called "Valdres sparagmite," a Caledonian flysch similar to the sparagmite.

The formation moved intermittently as a nappe on the underlying mass. This was accompanied by minor discontinuous movements along bedding planes, with the shale zones serving as a lubricant. Simultaneously, the rock itself underwent a continuous shearing movement, which was so pronounced that, in the axes are normal to the plane. Everywhere the long axes are oriented in a northwest-southeast direction, the direction of the main transport, which is marked by a pronounced lineation. This orientation of the pebbles agrees with what was found by Strand (1945, p. 17), and by Kvale (1945, p. 193).

## DEFORMATION OF THE CONGLOMERATES

### DEFINITIONS

In accordance with Cloos (1947, p. 853) the following designations have been used: The spherical pebbles have a radius r; after deformation the ellipsoids have semiaxes, a, b, and c. The tectonic axes *a*, *b*, and *c* are oriented as defined by Sander and many other authors. It should be emphasized that the direction of the lineation is parallel to the direction of the main transport. The same relation was found by Kvale (1945, p. 204), who considered the *a* axis to be parallel to the lineation.

### THE MEASUREMENTS

It is very difficult to find pebbles in which all the three ellipsoidal axes can be measured, as emphasized by Strand (1945, p. 18). In most cases the pebble has to be cut out of the rock so as to secure a reliable measurement of all axes. In only three localities, where the planes of schistosity are extensively exposed, were more than 10 pebbles measured. In most cases it has been possible to measure with an exactness up to 1 mm., but not always, as shown by table 1.

During six years of field work I have measured 118 pebbles (table 1). In addition, ten measurements from similar rocks are taken from Strand (1938, p. 44; 1945, p. 19). The two localities of Strand are found within the "Valdres Sparagmite," a flysch horizon situated between two nappes of igneous rocks overlying the Sparagmite formation.

### THE MODE OF DEFORMATION

Before treating the measured properties, it is necessary to consider the type of deformation.

The deformed rocks, described by Cloos (1947), consist of shales and limestones. These rocks were shear-folded by the orogenic movement, so that the original thickness of the beds increased. The Sparagmites, on the contrary, have been thrust without folding. The shear movement which deformed the conglomerates (and the rocks) could have acted along one or more shear planes. In a quartzite from western Norway, Kvale (1945, p. 199) found three planes of shear, one macroscopic and two microscopic. The latter planes were indicated by the orientation of the micas.

The deformed Sparagmites have one macroscopic *s* plane—the plane of schistosity. This plane is defined by a marked cleavage and by the parallel arrangement of elongated quartz and feldspar grains and of the faintly wavy zones of muscovite. A study of thin sections from the Rondane area seems to indicate that the muscovite has only one preferred orientation, which is parallel to the elongated quartz and feldspar grains and parallel to the plane of schistosity. No other macroscopic or microscopic structure is observed.

These facts suggest that the conglomerates were deformed by shear parallel to the plane of schistosity. The general case is then a triaxial deformation by inhomogeneous shear. But, in order to calculate the deformation, some simplifying assumptions are necessary.

First, the shear is assumed to be homogeneous within each layer of conglomerate. The shear movement along *a* is no doubt the principal movement, and shear in the *b* direction plays a subordinate role. Therefore, the shear may be assumed to be, as a first approximation, a pure shear in the *a* direction.

Many authors have discussed this type of deformation, and it has many names: "shearing motion (or scission)" by Becker (1893, p. 24), "Scherbewe-

gung" by Sander (1930, p. 15), "simple shear" by Nadai (1931, p. 294), and "einscharige Gleitung" by Schmidt (1932, p. 45). Following Becker, the author prefers "shearing motion."

Becker (1893) was the first to give mathematical expressions to the different types of rock deformation which might occur. But he did not deduce the equations of the shearing motion. This is done by Nadai (1931, pp. 295–297).

The shearing motion may be defined in the following way:

The rock is cut up into infinitely thin

TABLE 1

DIAMETERS OF THE MEASURED PEBBLES*

| Locality | Pebble Dimensions (Cm.) | | | d |
|---|---|---|---|---|
| 1 | 3.5 | 2.8 | 1.8 | 2.60 |
| | 8.3 | 7 | 3 | 5.59 |
| | 8.8 | 6.7 | 2.8 | 5.49 |
| 2 | 9 | 6 | 4.5 | 6.25 |
| | 7 | 7 | 4 | 5.81 |
| | 6.5 | 6 | 3 | 4.90 |
| | 6 | 3 | 2.3 | 3.46 |
| | 7 | 6 | 4.4 | 5.70 |
| | 8.3 | 4.8 | 3 | 4.93 |
| | 4.5 | 3.3 | 2.1 | 3.15 |
| | 6.5 | 4.5 | 2.3 | 4.07 |
| | 7.5 | 4.7 | 2.8 | 4.63 |
| | 6.5 | 4.8 | 3.8 | 4.91 |
| | 14 | 9.5 | 2.4 | 6.84 |
| | 8.5 | 6.2 | 3.5 | 5.69 |
| | 7.5 | 3 | 2 | 3.56 |
| | 9.5 | 7 | 3.5 | 6.15 |
| 3 | 18 | 9.7 | 3.5 | 8.49 |
| | 23 | 10 | 3 | 8.84 |
| | 12 | 5 | 2 | 4.94 |
| | 14 | 10.5 | 3.5 | 8.01 |
| | 12 | 5 | 1.5 | 4.49 |
| | 9 | 6 | 2.2 | 4.90 |
| | 7.5 | 5 | 1.8 | 4.07 |
| | 17 | 10 | 2 | 7.00 |
| | 6.3 | 4.3 | 1.1 | 3.10 |
| | 13.5 | 6 | 2 | 5.48 |
| | 8 | 4.5 | 2.2 | 4.30 |
| | 4.8 | 2.8 | 1.1 | 2.46 |
| | 5.3 | 4 | 1.5 | 3.17 |
| | 10 | 7.5 | 1.7 | 5.03 |
| | 11.2 | 5.2 | 1.4 | 4.34 |
| | 21 | 11 | 2.5 | 8.32 |
| | 8.5 | 5.3 | 1.4 | 3.98 |
| | 11 | 6 | 2 | 5.10 |
| | 17 | 8.5 | 3 | 7.56 |
| | 6.7 | 4.7 | 1.8 | 3.84 |
| | 6 | 4.5 | 2.3 | 3.96 |
| | 13.5 | 9 | 2 | 6.24 |
| | 10 | 6 | 2.5 | 5.32 |
| | 24.5 | 11.2 | 3.6 | 9.95 |
| | 9 | 6 | 2.5 | 5.13 |
| | 12.5 | 7 | 2.6 | 6.10 |
| | 8 | 6 | 2.2 | 4.73 |
| | 9.5 | 4.8 | 1.8 | 4.35 |
| | 8.3 | 6 | 2.4 | 4.92 |
| | 8 | 4.4 | 2.0 | 4.13 |
| | 17 | 8 | 2.5 | 6.98 |
| | 13 | 8.8 | 3 | 7.00 |
| | 9.5 | 3.5 | 1.2 | 3.42 |
| | 7.3 | 4.5 | 1.7 | 3.82 |
| | 8 | 5 | 1.7 | 4.08 |
| | 17.5 | 14.5 | 4 | 10.05 |
| | 5.5 | 3.7 | 2 | 3.44 |
| | 12.5 | 9 | 2.1 | 6.18 |
| | 11 | 7 | 1 | 4.26 |
| | 5.5 | 4 | 1.1 | 2.90 |
| | 9 | 5.5 | 1.8 | 4.47 |
| | 8.5 | 6 | 1.3 | 4.05 |
| | 16 | 10 | 1.8 | 6.65 |
| | 6.2 | 4.5 | 1.3 | 3.31 |
| | 10 | 5 | 2 | 4.65 |
| | 15.5 | 12 | 2.5 | 7.75 |
| | 17 | 8.5 | 2.3 | 6.92 |
| | 6.5 | 4.5 | 2.5 | 4.18 |
| | 8.5 | 6 | 1.7 | 4.43 |
| | 14 | 8 | 3 | 6.96 |
| | 8.4 | 4 | 1.8 | 3.92 |
| | 8.5 | 6.5 | 2 | 4.82 |
| | 7 | 6 | 2.3 | 4.59 |
| | 7 | 5 | 2.5 | 4.44 |
| | 7 | 6 | 1.5 | 3.98 |
| | 5 | 5 | 1.6 | 3.42 |
| | 10 | 7 | 2.5 | 5.60 |
| | 8.6 | 6 | 2 | 4.69 |
| | 10 | 8 | 2 | 5.44 |
| 4 | 22 | 9 | 4.5 | 9.60 |
| | 14 | 9 | 4 | 7.96 |
| | 19 | 10.5 | 3.4 | 8.78 |
| 5 | 6 | 2.5 | 1 | 2.47 |
| | 6 | 3 | 1.2 | 2.78 |
| 6 | 12 | 5 | 2 | 4.94 |
| | 8 | 5 | 1 | 3.42 |
| | 8 | 6 | 1.5 | 4.16 |
| | 9 | 6 | 2.5 | 5.16 |
| | 8 | 4 | 2 | 4.00 |
| 7 | 20 | 6 | 2 | 6.22 |
| 8 | 6 | 2.5 | 1.2 | 2.62 |
| | 6 | 3 | 1 | 2.62 |
| 9 | 30 | 6 | 0.5 | 4.49 |
| | 20 | 5 | 0.3 | 3.11 |
| | 20 | 5 | 0.4 | 3.42 |
| | 30 | 5.5 | 0.4 | 4.05 |
| | 20 | 4 | 0.4 | 3.18 |
| | 22 | 5.5 | 0.7 | 4.40 |
| | 25 | 6 | 0.4 | 3.92 |
| 10 | 12 | 3.5 | 1.3 | 3.80 |
| | 12 | 4 | 1.3 | 3.97 |
| | 5 | 3.5 | 1.1 | 2.68 |
| | 7.5 | 3.8 | 1.2 | 3.25 |
| | 9 | 3 | 1.1 | 3.10 |
| | 5 | 3.3 | 0.8 | 2.36 |
| | 6.8 | 2.5 | 0.9 | 2.48 |
| | 9 | 2.4 | 0.9 | 2.69 |
| | 8 | 3.6 | 0.8 | 2.85 |
| | 5.8 | 3.7 | 1.0 | 2.78 |
| | 8 | 3.7 | 1.1 | 3.20 |
| | 4.5 | 2 | 1.2 | 2.21 |
| | 5.5 | 3.4 | 0.9 | 2.56 |
| | 6.5 | 4 | 3.4 | 4.46 |
| | 5.5 | 2.8 | 1.1 | 2.58 |
| | 4.9 | 2.9 | 1.2 | 2.58 |
| | 8.2 | 3.5 | 1.4 | 3.43 |
| | 6.6 | 5.2 | 1.0 | 3.44 |
| | 6.7 | 3.2 | 0.7 | 2.47 |
| | 8.2 | 1.9 | 1 | 2.50 |
| | 10 | 2.4 | 1.8 | 3.51 |
| 11 | 30 | 3.3 | 0.3 | 3.10 |
| 12 | 45 | 10 | 3 | 11.08 |
| | 20 | 9 | 2 | 7.12 |
| | 32 | 11 | 7 | 13.50 |
| | 24 | 7 | 3 | 7.96 |
| 13 | 39 | 12 | 7 | 14.89 |
| | 37 | 9 | 6 | 12.61 |
| | 28 | 9 | 6 | 11.50 |
| | 18 | 8 | 5 | 8.95 |
| | 15 | 4 | 1 | 3.92 |
| | 13 | 4 | 1 | 3.74 |

* The original diameters (*d*) were calculated from eq. (7), p. 481. The localities are tabulated under figure 1.

lamellae parallel to the plane of schistosity, and each lamella is moved a certain distance in relation to the underlying lamella. Then the "amount of shear" (Becker, 1893, p. 25) is the distance *divided by* the thickness of the lamellae. The direction of movement is the tectonic *a* axis of Sander (1930), and the *c* axis is perpendicular to the plane of schistosity.

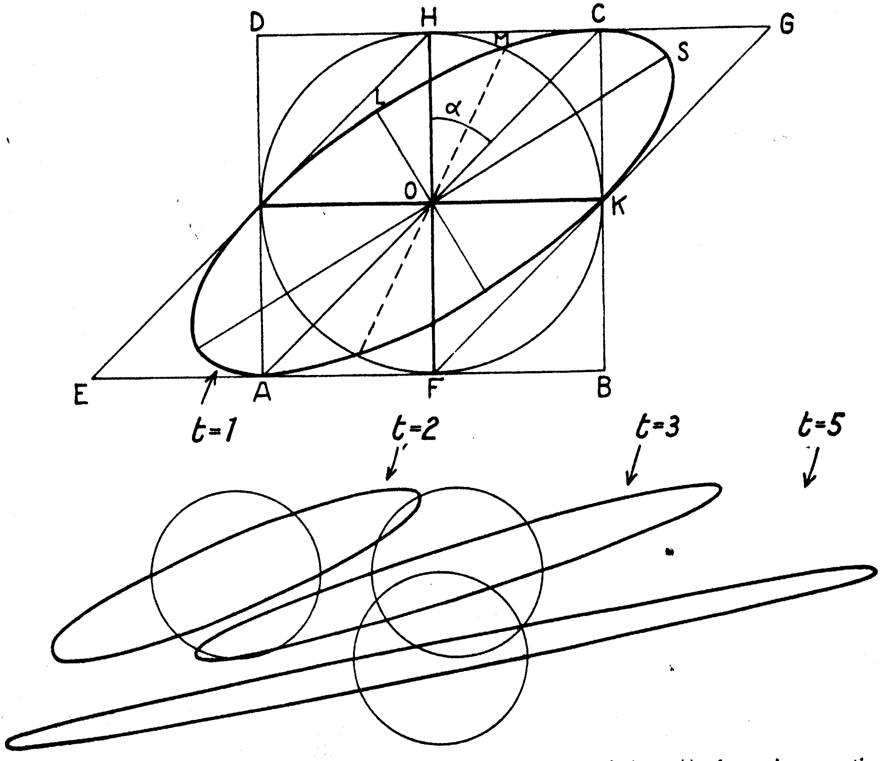



FIG. 2.—Deformation by shearing motion with varying amounts of shear ($t$), drawn in *ac* sections. *Above:* With $t = 1$ the square *ABCD* is transformed into the parallelogram *EFGH* (*OK* being plane of relative movement), and the circle with radius *OK* is transformed into an ellipse with semiaxes *OS* and *OL*. *Below:* Deformations with amount of shear $t = 2$, 3, and 5.

Figure 2 shows the deformation of a sphere by increasing amounts of shear ($t$). The sphere with radius *OK* becomes an ellipsoid with semiaxes a = *OS* and c = *OL*, while b (vertical to the paper) remains constant = *OK*. With increasing amount of shear the ellipsoid is increasingly elongated and flattened.

If the radius *OK* = 1, the equation of the circle is

$$x^2 + y^2 = 1\,. \qquad (1)$$

By deformation, the circle is altered to a "strain ellipse" (Nadai, 1931, p. 295):

$$x^2 - 2txy + (t^2 + 1)\,y^2 = 1\,. \qquad (2)$$

By rotating the co-ordinate system, the formula of the ellipse becomes

$$\frac{x^2}{a^2} + \frac{y^2}{c^2} = 1 \qquad (3)$$

where

$$\left.\begin{matrix}a\\c\end{matrix}\right\} = \sqrt{\tfrac{1}{2}(2+t^2 \pm t\sqrt{t^2+4})} \quad (4)$$

and $a \cdot c = 1$.

The amount of shear ($t$) is easily defined by tan $\alpha = t$ (fig. 2), or from equation (4) by:

$$t = \sqrt{a^2+c^2-2}\,. \quad (5)$$

CALCULATION OF THE DEFORMATION

We assume that a spherical pebble is deformed. The volume remains constant:

$$\tfrac{4}{3}\pi r^3 = \tfrac{4}{3}\pi abc\,, \quad (6)$$

$$r = \sqrt[3]{abc} \quad (7)$$

or, when r = b,

$$r = \sqrt{ac}\,, \quad \text{or} \quad r^2 = ac\,. \quad (8)$$

Then

$$\frac{a}{r} = \frac{r}{c}\,. \quad (9)$$

This ratio, given as a percentage (100 a/r = 100 r/c), may be taken as the deformation ratio of the a and c axes.

But it is easy to find that b always deviates more or less from $\sqrt{ac}$. Thus the b axes also undergo a deformation. Then r is defined by $\sqrt[3]{abc}$. Still it is advantageous to use 100 a/r and 100 r/c as the deformation ratios of the axes a and c. Now they differ from each other; and the more they differ, the larger the deformation of b. The latter deformation is given by 100(b − r)/r, as defined by Cloos (1947, p. 863).

This way of calculating differs slightly from that of Cloos (1947, p. 863). He gives all three components of deformation by the difference from the original radius, in percentage of this radius; for example, for the c axis: 100 (c − r)/r.

By this procedure Cloos gets various values for the deformation in *a* and *c*, even if the deformation is of the ideal type, namely, with no distortion in *b*. For example, if the dimensions in centimeters are 100, 10, and 1. Then r = $\sqrt[3]{1000}$ = 10. Following Cloos, the deformation in *a* is

$$\frac{100-10}{10}\cdot 100 = 900 \text{ per cent}\,,$$

and in *c:*

$$\frac{1-10}{10}\cdot 100 = -90 \text{ per cent}\,.$$

By the calculation suggested above the corresponding properties are:

$$\frac{100}{10}\cdot 100 = 1{,}000 \text{ per cent}\,,$$

$$\frac{10}{1}\cdot 100 = 1{,}000 \text{ per cent}\,.$$

This property, 1,000 per cent, expresses the deformation of the rock.

In figure 3 the deformation of a is plotted against the corresponding deformation of c. Points on the broken line have equal deformation ratios of a and c, and therefore they mark the pebbles which show no distortion of b. Most of the points lie above the line, indicating a higher deformation of c than of a. This means that b has been extended. Another expression for this relation is obtained by a plot analogous with figure 12 of Cloos (1947, p. 889), where the deformation of b is plotted against the deformation of a (fig. 4).

Figure 4 shows that the pebbles under consideration exhibit a much greater deformation in *b* than did the oöids of Cloos. In addition to an extension in *b*, reduction in *b* is not rare. As already seen from figure 3, deformation in *a* (and *c*) is also, in general, much more intense than that observed by Cloos.

There is another way of characterizing

the deformation, defined by the amount of shear. For example, in the above-mentioned case, the deformation is 1,000 per cent, but the amount of shear is about 10, according to equation (5). The property 10 means that, if the conglomeratic bed, 1 meter thick, is deformed homogeneously, the upper part of it has moved 10 meters in relation to the lower part. Thus the amount of shear gives a much clearer picture of the degree of deformation than does the deformation percentage, and it is called the "gradient of deformation" in the following discussion.

An ellipsoidal pebble with semiaxes a, b, and c may be characterized by the double ratio a : b : c. In the form

$$\frac{a}{c} : \frac{b}{c} : 1$$

the ratios are independent of the absolute size, and each pebble may have its

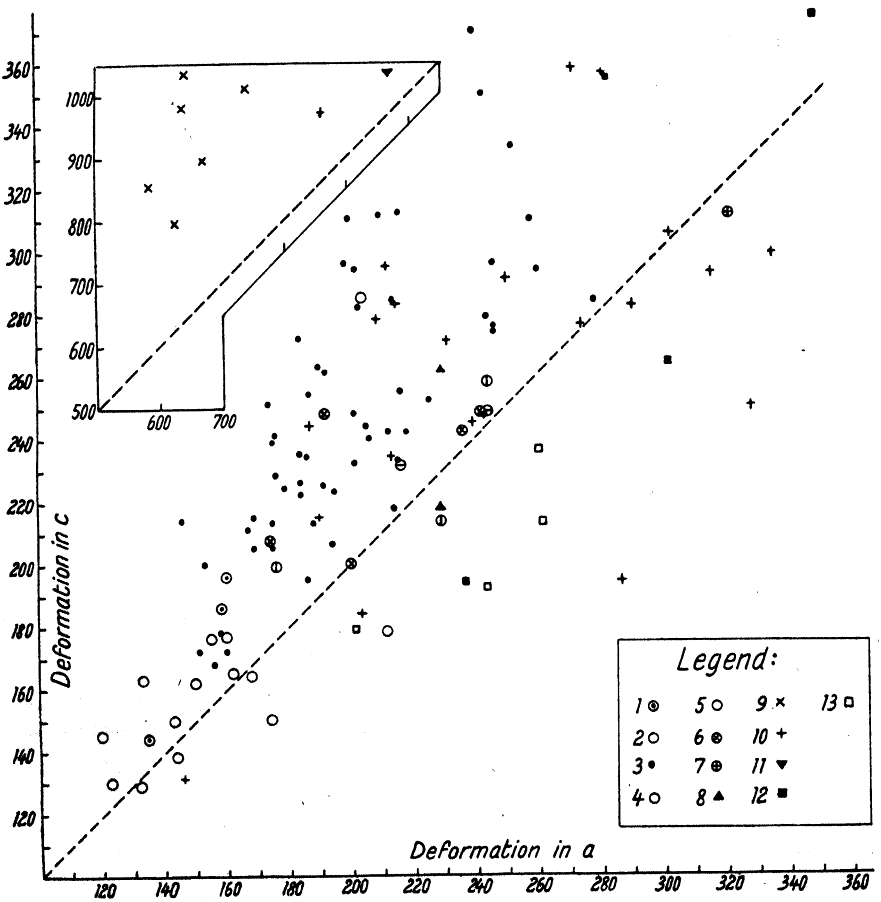



FIG. 3.—Deformation of a axes in relation to deformation of c axes, given by the expression 100 a/r and 100 r/c, viz. The numbers in the legend refer to the localities, as tabulated under fig. 1. *Inserted:* Highly deformed conglomerates.

deformation marked by the values a/c and b/c. Equation (4) is deduced from the deformation of a unit sphere (r = 1). Then b = r = 1, and the axial ratios are a : 1 : c or

$$\frac{a}{c} : \frac{1}{c} : 1 .$$

By giving *t* the values 1, 2, 3, etc., in equation (4), the values a/c and 1/c may be given as a function of *t*. These values give the two full-line curves of figure 5, one for a/c and one for b/c. In this diagram the deformed pebble may be easily plotted as a point by the following procedure: From a = 4, b = 2, and c = 1, the values a/c = 4 and b/c = 2 are calculated. The point on the curve for b/c with an ordinate equal to 2 is found, and vertically above it, with an ordinate equal to 4, the point is plotted. This point will lie on the curve for a/c, if the pebble is deformed by a pure shearing motion. Thus figure 5 indicates in a new way that the deformation of most of the measured pebbles deviates from pure shearing motion. But the plots, which fall near the a/c curve, directly define a gradient of deformation.

PROBLEMS OF THE DEFORMATION

The plots of figures 3, 4, and 5 show that the deformation of the pebbles was not solely a shearing motion. The movement, therefore, may be of a more complicated type, and, in addition, some other factors may influence the result. Various facts suggest such assumptions.

First, the present orientation of the pebbles does not agree with the theory of shearing motion, because then the long axis of the ellipsoidal pebble should form a certain angle with the plane of schistosity. On figure 2 this angle is (90 − α)°. In mildly deformed conglomerates the angle should be easily observed. However, the long axis is lying in the plane of schistosity, or, in highly deformed conglomerates, forms a small angle with the plane in certain localities. The fact that the long axis, a, mostly

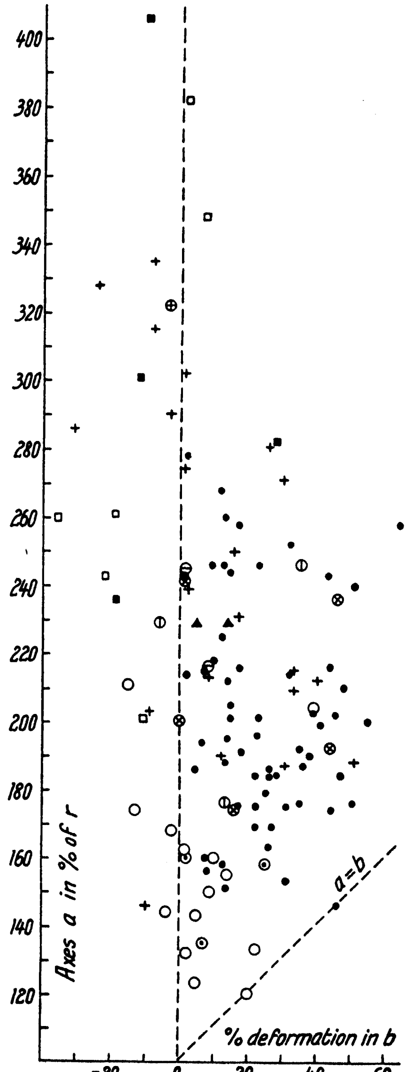



FIG. 4.—Deformation of b axes (given by 100 [b − r]/r) in relation to deformation of a axes, given 100 a/r as in fig. 3.

coincides with the tectonic axis *a*, may be explained either as being caused by an external rotation or by another type of deformation than shear on the plane of schistosity. I consider the second possibility as improbable, and I suggest that the whole pebble has been rotated, simultaneously with the shear deformation.

By inhomogeneous strain the pebbles should get a shape deviating from that of an ellipsoid. In *ac* and *bc* sections the pebbles are observed to have outlines which are scarcely exact ellipses. Even a flat ellipse is rounded at the ends, but the pebbles are commonly sharp. It is, however, difficult to get a quantitative

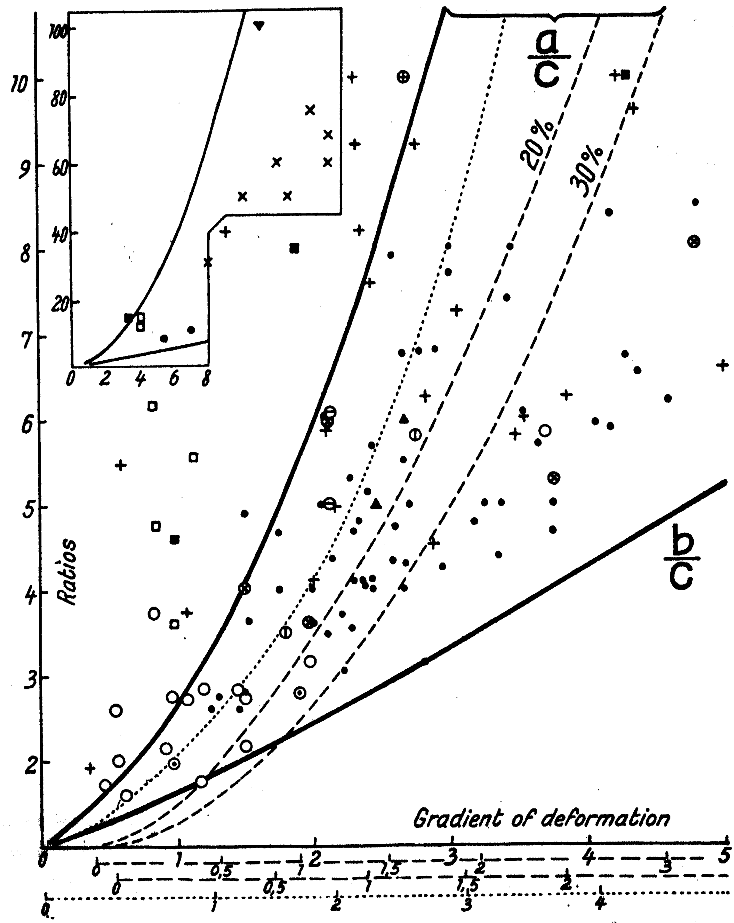



FIG. 5.—Deformation of pebbles by plots of the axial ratios a/c : b/c : 1 in relation to the gradient of deformation. The full-line curves give the values of a/c and b/c for a varying gradient by shearing motion. The values are calculated from equation (4). A pebble with axes a, b, and c is plotted as a point by the ratios a/c : b/c : 1. The value b/c is found on the curve for b/c, then the point has the same abscissa, and the ordinate a/c. If the point falls near the full-line curve for a/c, the deformation is a shearing motion, and the gradient of deformation may be read. Curves for a/c for shearing motion combined with elongation in b (*broken lines*) and for nonspherical shape of the original pebble (*dotted line*) has been constructed (see text, pp. 485 and 486). *Inserted:* Plots of highly deformed pebbles. The same designations are used. Only the shearing-motion curves for b/c and a/c are drawn.

measure of these deviations. The two major problems of the deformation are alteration of the b axes and the primary shape of the pebbles. These problems are discussed at some length below. They are considered independently.

### DISTORTION OF THE *b* AXES

If we assume that the deformed pebbles originally had a spherical shape, the present shape should give a true account of the deformation process. Accordingly, the shape should give quantitative expressions for the deformation in *b*.

Cloos pays considerable attention to the "distortion parallel fold axes" (pp. 886–892). He specially emphasizes the importance of elongation in *b* in arcuate orogens.

The Caledonian orogen forms an arc just in the area from which the Sparagmites have been thrust. Regional maps show that the orogen alters its direction a certain angle in the course of 300 km. The angle may be estimated to be 10°–20°. The exact elongation may then be calculated, if the necessary properties are determined. Let us assume that the arcuation is 10° over a width of 300 km.

In order to calculate the pebble deformation in a Sparagmite nappe with dimensions 300 × 300 km., some further simplifications must be made. If the pebbles have an external rotation simultaneously with the shearing motion, the calculation of the corresponding strain ellipse becomes very intricate. Therefore, the rotation is considered to occur after the shearing motion. Then the b axes are elongated, and the c axes are reduced, while the a axes remain unchanged.

In the front of the thrust masses the increase of the nappe parallel to the b axes is sin 10° · 300 km. = 52 km., or nearly a 20 per cent increase. A corresponding reduction occurs to the c axes.

From figure 4 the deformation in *b* may be read directly. But in order to express deformation in *a* and *c* by the easily understandable "gradient of deformation," the deformation must be considered as shearing motion combined with irrotational homogeneous strain parallel to the b axis.

A pebble with a radius r is deformed by shearing motion to an ellipsoid with semiaxes a, b = r, and c. The elongation in b of, say, 20 per cent changes the dimensions to a, 1.2 b, and c/1.2. Instead of

$$\frac{a}{c} : \frac{b}{c} : 1 ,$$

the ratios are

$$1.2\frac{a}{c} : (1.2)^2\frac{b}{c} : 1 .$$

From the latter ratios a number of values for the varying gradient *t* may be calculated (eq. [4]). These values define two curves.

There are two ways of drawing these curves for comparison with the ordinary shearing motion. Either the two curves may be drawn in figure 5 with the same unit of gradient as the shearing motion, or the already existing curve for b/c may be used as a reference curve, as explained for the plotting of measurements. For a certain *t* the ordinate $(1.2)^2(b/c)$ is found on the curve for b/c, and vertically above this point the ordinate $1.2(a/b)$ is plotted. Points for various *t* values define a curve. By this procedure the deformation with elongation in b is demonstrated by a new curve, which has other units for the gradient *t* on its abscissa axis.

The latter way of establishing curves is followed. On the new curves (broken curves on fig. 5) we should find all the points, which represent pebbles deformed by shearing motion with elongation in *b* equal to 20 per cent and 30 per cent, and a varying gradient of deformation. By extrapolation the extension in *b* and the

gradient of every observation may be read from figure 5.

### THE ORIGINAL PEBBLE SHAPE

No conglomerates have pebbles with a shape which mathematically is a sphere. Undeformed quartz conglomerates, similar to the described ones, are found along the southeastern border of the Sparagmite formation. The pebbles of these conglomerates have a shape which varies to a certain degree. All the pebbles are well rounded, and most of them are spherical or nearly so, but some pebbles are ellipsoidal, with one long axis in the bedding plane, or they are flat in this plane, that is, they are rotational ellipsoids with the rotation axis (short axis) vertical to the bedding plane. This latter shape is considered as the most typical of deviations from a spherical shape. The author has tried, therefore, to calculate the deformation of such an ellipsoid.

If the ellipsoidal pebble has the semi-axes a′, a′ = b′, and c′, its equation is

$$\frac{x^2}{a'^2}+\frac{y^2}{a'^2}+\frac{z^2}{c'^2}=1\ ,\quad \text{where } a'>c'\ . \tag{10}$$

In the ac section the corresponding ellipse,

$$\frac{x^2}{a'^2}+\frac{z^2}{c'^2}=1\ , \tag{11}$$

is transformed by shearing motion into another ellipse:

$$c'^2x^2-2c'^2txz+(c'^2t+a'^2)\,z^2=a'^2c'^2\ . \tag{12}$$

By rotation of the co-ordinate system, the formula of the ellipse becomes

$$\frac{x^2}{a^2}+\frac{z^2}{c^2}=1\ , \tag{13}$$

where

$$\left.\begin{matrix}a\\c\end{matrix}\right\}=\sqrt{\tfrac{1}{2}\left(a'^2+c'^2\,(t^2+1)\pm\sqrt{(a'^2+c'^2\,[t^2-1])^2+4t^2c'^4}\right)}\ . \tag{14}$$

We choose a′ = b′ = 1, c′ = $\frac{3}{4}$, as probable values for an ellipsoidal pebble. This shape is one of the most extreme deviations from the sphere. The deformed pebble has the axial ratios (b′ = 1 is constant) a : 1 : c. On the usual form:

$$\frac{a}{c}:\frac{1}{c}:1\ .$$

By giving $t$ the values 1, 2, 3, etc., we obtain a set of values for a/c and for 1/c which define two curves. In order to compare the plotted observations of figure 5 with this new case, the two curves are substituted by one new curve in figure 5, using the old b/c curve as a reference curve, as described for the curves showing the extension in b.

The pebbles which have their points lying on the dotted curve of figure 5 may accordingly originate from deformation by shearing motion, if the original pebble was a flat rotational ellipsoid with semi-axes 1, 1, and $\frac{3}{4}$. And, with the original pebble shape lying between this ellipsoid and the sphere, the points of the deformed pebbles should lie statistically between the full line and the dotted a/c curve, if the deformation type was shearing motion. However, relatively few points lie between the mentioned curves, and this fact proves that the deviation from the spherical shape of the original pebbles is not a factor of primary importance in relation to extension in *b*.

### CONCLUSIONS

The facts and calculations presented above show that the deformation of the measured pebbles is of a very complex nature. The conclusions cannot be as quantitative as intended, and they may

be subject to revision when more material is collected. But some qualitative results are obtained, and some semiquantitative estimates are given.

1. Shearing motion is considered as the main element of the deformation, but figure 5 indicates that, in addition, a deformation in b is evident. The deformation varies from increase (up to 30 per cent) to decrease (to about 30 per cent). The deviation from a sphere in the original pebble shape does not influence severely the significance of the figures but probably causes the variation within each locality.

2. The deformation in a and c increases from the border of the orogen toward the central zone. For the different localities the gradient of deformation may be read as follows: Nos. 1 and 2: gradient below 1. No. 3: mostly between 1 and 1.5; a few above 2. Nos. 4, 5, 6, 7, and 8: between 1 and 2. No. 10: mostly between 2 and 3. No. 9: probably from 6 to 10. No. 11: 10. Nos. 12 and 13: between 2 and 4.

3. The mean extension in b may be estimated as follows for the different localities: Nos. 1 and 2: uncertain. No. 3: 20 per cent. Nos. 4, 5, 6, 7, and 8: between 10 and 20 per cent. No. 10: 10 per cent. No. 9: between 20 and 30 per cent. No. 11: 0 per cent. Nos. 12 and 13: mostly a pronounced reduction in b (20 per cent).

Thus the estimates show that the extension in b decreases from the border toward the central zone. In highly deformed conglomerates even a pronounced reduction occurs.

The decrease of the extension may be connected with the arcuated orogen.

ACKNOWLEDGMENTS.—The field observations for the present paper were collected during geological mapping for the Geological Survey of Norway. The author is indebted to Professors F. J. Pettijohn and R. B. Balk for corrections and suggestions.

## REFERENCES CITED

BECKER, G. F. (1893) Finite homogeneous strain, flow, and rupture of rocks: Geol. Soc. America Bull. 4, pp. 13–90.

CLOOS, ERNST (1947) Oölite deformation in the South Mountain fold, Maryland: Geol. Soc. America Bull. 58, pp. 843–918.

KVALE, ANDERS (1945) Petrofabric analysis of a quartzite from the Bergsdalen quadrangle, western Norway: Norsk geol. tidsskr., vol. 25, pp. 193–215.

NADAI, A. (1931) Plasticity, New York, McGraw Hill Book Co., Inc.

SANDER, BRUNO (1930). Gefügekunde der Gesteine, Wien, Julius Springer.

SCHMIDT, WALTER (1932) Tektonik und Verformungslehre, Berlin, Borntraeger.

STRAND, TRYGVE (1938) Nordre Etnedal. Beskrivelse til det geologiske gradteigskart: Norges geol. undersökelse, no. 152.

——— (1945) Structural petrology of the Bygdin conglomerate: Norsk geol. tidsskr., vol. 24, pp. 14–31.

# GEOLOGICAL NOTES

## A STEREOGRAPHIC CALCULATOR[1]

ROBERT E. WALLACE
State College of Washington, Pullman, Washington

### DESCRIPTION OF INSTRUMENT

The stereographic calculator here described represents essentially a combination in one physical unit of the stereographic net, Penfield paper, and a stereographic scale, thus eliminating certain mechanical difficulties in manipulating the units separately. The combination results in a sturdy, lightweight instrument, like a slide rule in action, with which problems solvable by the stereographic projection can be done easily and rapidly either in the field or in the office.

The instrument (fig. 1) is composed of three main parts: (1) a stereographic net on a circular disk of metal or cardboard; (2) a graduated circle on a circular disk of transparent plastic, the front of which is frosted to take pencil; (3) a circular metal housing in which the disks carrying the stereographic net and graduated circle can revolve freely. This metal housing is made in two parts, a back and a front, so that it may be opened and the disks inserted. A circular portion of the back half is cut away so that the disk with the stereographic net can be revolved with the fingers. The front half is cut away so as to expose the net and graduated circle in a semicircular opening and produce a straight edge which bisects the stereographic net. As much of the remainder of the front can be cut away as will produce better visiblity of the net and yet retain the strength of the straight edge as well as of the metal housing itself. On the straight-edge portion of the front a stereographic scale is inscribed.

In the instrument prepared as a model, the housing was made of magnesium, although that metal proved somewhat unsatisfactory, in that it marked the frosted disk relatively easily. A stereographic net, 10 cm. in diameter, on cardboard backing, prepared by Fisher and distributed by Ward's Natural Science Establishment, Inc., of Rochester, New York, was used and cut into disk form to make the rear disk. Ribs were glued to the back of the disk, so that torque could be applied to rotate the disk easily. The front disk was made of ordinary transparent celluloid; the graduated circle was inscribed by hand with a stylus, and the front was frosted by a few minutes' grinding with fine-grained abrasive on a glass surface.

### USE OF INSTRUMENT

Many simple problems, such as calculating the apparent dip of a bed in any vertical section cutting the bed (fig. 2) can be solved merely by rotating the two disks and scale into the proper positions with respect to each other and reading the answer directly. Other problems require marking a point or line on the frosted surface of the disk containing the graduated circle and then rotating the disks and scale to a position where the answer may be read. The method for solving structural problems as described by Bucher (1944) will suggest clearly the procedure for using the present instrument.

Although any problem involving stereographic projection can be done on this instrument, the solution of certain problems can be further simplified by using disks with different combinations of nets inscribed on them. For example, if one has numerous problems involving the determination of the trend and plunge of the line of intersection of two planes, a combination of two disks, each inscribed with nothing but great-circle

[1] Manuscript received April 29, 1948.

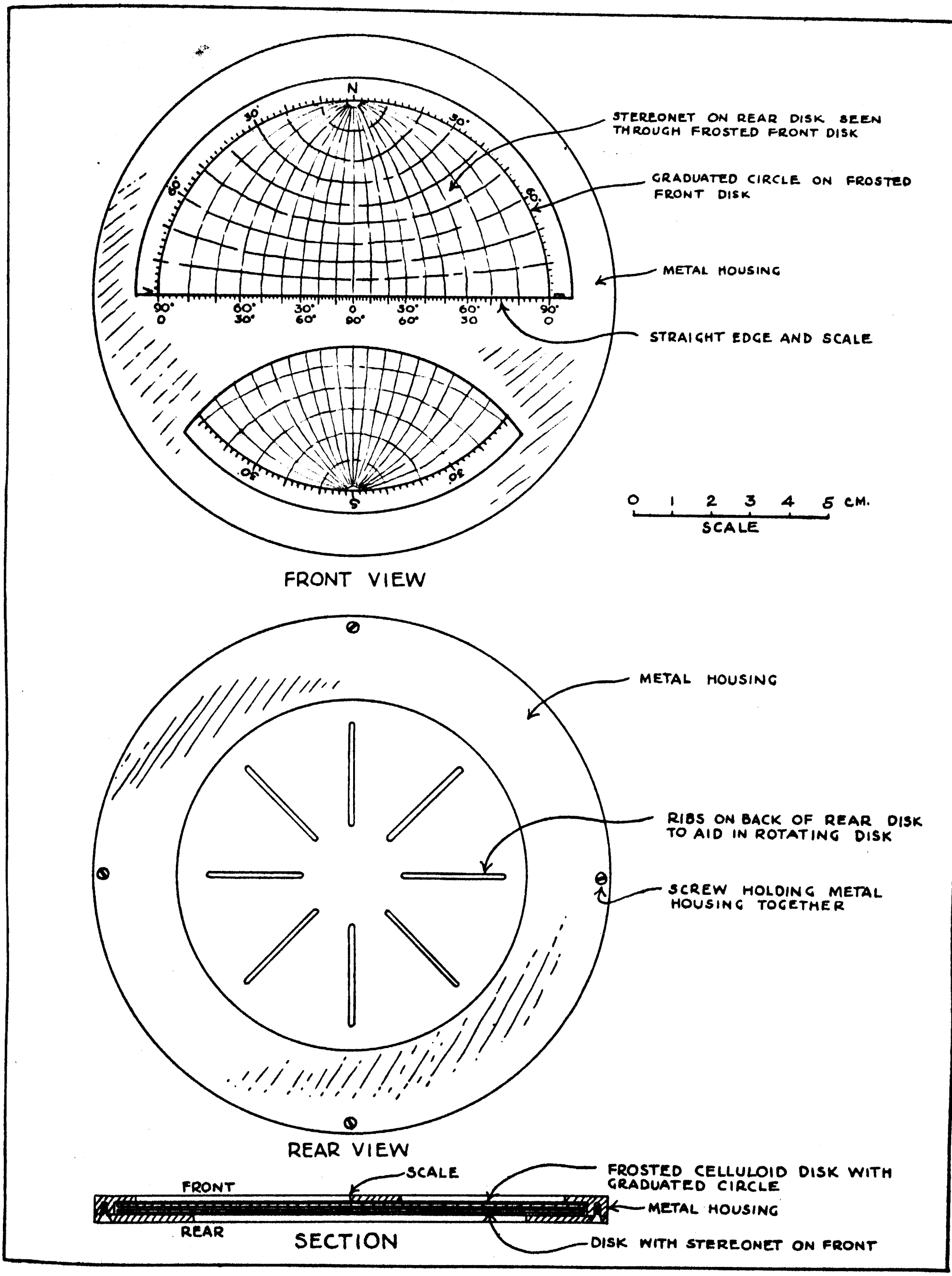



FIG. 1.—Diagrammatic sketch of stereographic calculator, showing construction of instrument

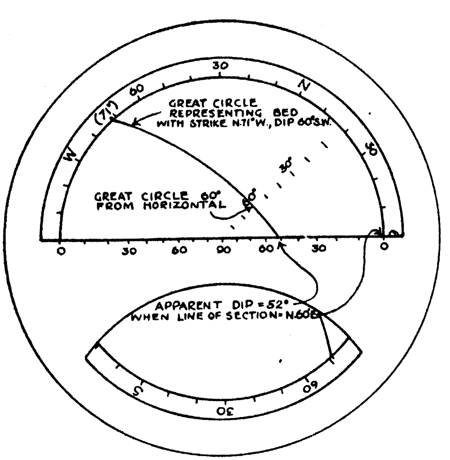



FIG. 2.—Example showing direct solution of apparent dip problem, where bed strikes N. 71° W., dips 60° SW., and strike of the line of vertical section is N. 60° E.

projections and degree divisions around the circumference of the projection, makes solution a matter of reading answers directly after proper rotation of the disks. Inasmuch as the metal housing can be taken apart readily, any of a variety of disks can be substituted to simplify the solution of certain problems.

The instrument has proved very satisfactory as a sturdy field and office tool for calculation of structural problems. In the field the easy solution of problems makes it possible to realize relations that might otherwise be overlooked, and pertinent study is suggested.

## REFERENCES CITED

BUCHER, W. H. (1944) The stereographic projection, a handy tool for the practical geologist: Jour. Geology, vol. 52, pp. 191–212.

FISHER, D. J. (1941) A new projection protractor: Jour. Geology, vol. 49, pp. 292–323, 419–442.

# DETERMINATION OF SODIUM AND POTASSIUM IN SILICATE MINERALS AND ROCKS[1]

LARS LUND
University of Oslo

Several authors have described improvements of the Lawrence-Smith method for the determination of alkalies in silicates. Willard, Liggett, and Diehl (1942, p. 234) and Marvin and Woolaver (1945, p. 554) recommend hydrofluoric acid together with perchloric acid for disintegration of the sample. This solvent is also suggested by Lundell and Knowles (1927, p. 849). To remove excess hydrofluoric acid, Willard, Liggett, and Diehl (1942, p. 234) distil it off as siliconfluoride, whereas Marvin and Woolaver (1945, p. 554) fume it off by repeated evaporations with perchloric acid. To convert the perchlorates into oxides and chlorides, they use heating in a furnace at 550° C.

[1] Manuscript received February 11, 1948.

In both cases special precautions have to be taken to get rid of interfering elements. If phosphorus is present, the methods have to be modified. The procedures appear to require more skill and training and to take more time than the Lawrence-Smith method. The only step in the Lawrence-Smith procedure that needs great caution is the mixing of the sample with the flux and the transferring of it to the crucible. The mixing, however, can be done without any perceptible loss after a little training. The great advantage of the fluxing procedure is that it leaves the alkalies in solution, free from interfering elements, after but a few, simple operations. Small amounts of calcium and magnesium might contaminate the alkali chlorides and will effect the determination

of sodium if this is done by deducting it from the sum of the chlorides. By determining the sodium as $NaZn(UO_2)_3(CH_3COO)_9 \cdot 6H_2O$, however, this uncertainty can be eliminated.

Miller and Traves (1936, p. 1390) give a thorough discussion of the determination of sodium as the triple salt and give some experimental facts about the influence of calcium. Their results show that small amounts of calcium are of little or no consequence. After adding 1 gm. of $CaCl_2 \cdot 6H_2O$ to a solution from which 14.4 mg. of sodium chloride is to be precipitated, a double precipitation gives results only 0.1–0.2 per cent too high. (First precipitation was approximately 2 per cent too high.) The amount of calcium by the second precipitation will be of the same order of magnitude as in the case of the alkali chlorides after the Lawrence-Smith procedure has been followed.

During a short stay at the Department of Geology of the University of Chicago, I had the opportunity of making a few experiments investigating these problems.

The alkali content of several mineral and rock specimens was determined by the Lawrence-Smith method, and the potassium was determined as $K_2PtCl_6$. Instead of discarding the filtrate after the precipitation, the sodium was determined as the triple salt after the only interfering element, platinum, had been reduced by evaporation of the alcoholic filtrate in a conical flask on the waterbath. If, after evaporation to dryness, the reduction is not complete, a few milliliters of alcohol are added and evaporated as before. The reduction is complete when the yellow color of the platinum complex has disappeared. The reduced platinum occurs either as a powder or in thin flakes and is easily washed free of chlorine ions. Sodium was precipitated from the filtrate as the triple salt, by the procedure described by Kolthoff and Sandell (1945, p. 416).

B. Bruun and R. Higazy in the laboratory of the Geology Department made several determinations of sodium and potassium in rocks and minerals, with the Lawrence-Smith procedure as described by Washington (1930, p. 222). From the filtrates after the precipitation of potassium as chloroplatinate, the sodium was precipitated as triple salt. The results are given in table 1:

TABLE 1

PER CENT $Na_2O$

| Higazy's Determination | Triple Salt Determination |
|---|---|
| 1.32 | 1.21 |
| 11.03 | 10.78 |
| 2.98 | 2.98 |
| **Bruun's Determination** | **Triple Salt Determination** |
| 2.99 | 3.00 |
| 10.35 | 10.07 |
| 1.75 | 1.74 |
| 2.12 | 2.03 |

There is a tendency to get somewhat lower results by determination as triple salt. This must be ascribed to the fact that calcium and magnesium are not always completely removed before the sum of the alkali chlorides is determined, but this does not effect the determination of sodium when it is precipitated as triple salt.

This procedure may be especially useful when the amount of sodium is small compared with that of potassium.

ACKNOWLEDGMENTS.—I take pleasure in expressing my thanks to Dr. W. H. Newhouse, chairman of the Department of Geology at the University of Chicago, who placed all facilities of the department at my disposal and made this investigation possible. My thanks go to Dr. T. F. W. Barth, of the same department, for his never failing interest and encouragement.

## REFERENCES CITED

KOLTHOFF, I. M., and SANDELL, E. B. (1945) Textbook of quantitative inorganic analysis, New York, Macmillan Co.

LUNDELL, G. E. F., and KNOWLES, H. B. (1927) The analysis of soda-lime glass: Am. Ceram. Soc. Jour., vol. 10, p. 849.

MARVIN, G. B., and WOOLAVER, L. B. (1945) Determination of sodium and potassium in silicates. An improved method: Ind. and Eng. Chemistry, vol. 17, p. 554.

MILLER, C. C., and TRAVES, F. (1936) The determination of sodium and potassium in insoluble silicates: Jour. Chem. Soc. London, p. 1390.

WASHINGTON, H. S. (1930) The chemical analysis of rocks, New York, John Wiley & Sons, Inc.

WILLARD, H. H.; LIGGETT, L. M.; and DIEHL, H. (1942) Direct determination of potassium in silicate rock: Ind. and Eng. Chemistry, vol. 14, p. 234.


# SHALLOW-WATER ORIGIN OF RADIOLARITES IN SOUTHERN TURKEY[1]


S. W. TROMP

Fouad I University, Abassia (Cairo), Egypt


Radiolarites in Turkey and also in many other parts of the world are closely associated with basic volcanic rocks and manganese concretions and are shallow-water deposits connected with volcanic activity either continental or submarine. Fossil, abyssal radiolarite deposits are probably exceptional occurrences.

In the Malm (Upper Jurassic) of southern Turkey, particularly in the central Toros geanticline (Tromp, 1947, p. 367, Table II), a thick series of sediments occurs which has been described by Ortynski as "red beds" and by Lahn, Arni, Blumenthal, and others as "radiolarite series." It is composed mainly of red shales with many red and a few green chert beds, whose number increases upward in this section. Several of the red chert beds are rich in radiolarians, a fact which gave rise to the name "radiolarite series." Many of the chert beds are nonfossiliferous, however, and the name "red beds" would be preferable. But the name "radiolarites" is so generally used for all these red chert beds, both in Turkey and in other parts of the world, that we shall continue to use this latter name.

Most geologists consider the radiolarites an indication of deep-sea environment, as similar deposits are known in present-day oceans at depths of over 4,000 meters. However, most, if not all, radiolarite deposits in Turkey are shallow-water sediments. This can be proved as follows:

1. Red chert beds, with or without radiolarians, are a most common feature not only in the Jurassic radiolarite series but also at the border of large serpentine massives (mostly gabbroic rocks, metamorphosed later into serpentines). They are covered by basal conglomerates or other coarse sediments belonging to the basal portion of a younger overlapping series, indicating a shallow-water origin of the radiolarites. If those chert beds had been formed in an older period and did not belong to the new cycle of sedimentation, it would be remarkable that the erosion always preserved both the serpentines and their red cherty border zones.

2. The Jurassic series in southeastern Turkey is underlain by the so-called "flysch series," which is composed mainly of nonfossiliferous sandstones, many of which are very coarse.

3. At the end of the period of the radiolarite series, the central Toros geanticline rose above sea level, resulting in the Lower Cretaceous unconformity. This suggests shallow-water conditions at the end of the radiolarite period.

[1] Manuscript received March 16, 1948.

There is also another phenomenon which supports the assumption that radiolarites in Turkey are shallow-water deposits, explaining, at the same time, the origin of these radiolarites.

The radiolarites and red cherts in southern Turkey are always associated, either directly or indirectly, with basic intrusions or extrusions. First, at the end of, and perhaps even during, the Upper Jurassic, basic intrusions took place. Second, during the beginning of the deposition of the Upper Jurassic sediments, several of the Paleozoic serpentine massives were above sea level, mainly because of tilted block movements at the end of the Dogger. Because of erosion, much of the flysch series is missing, and the radiolarite series directly overlies the old serpentine cores (for example, near Egridir).

The influence of basic igneous rocks on radiolarite development may be explained as follows: If basic rocks are dissolved in the sea or if volcanic submarine exhalations and extrusions took place, the sea water would become very rich in silicic acid, which makes the milieu very favorable for the growth of radiolarians. This assumption is supported by various observations in other countries.

1. Recent plankton researches of Johnstone and others showed that, if the silicic acid content of sea water increases, the growth of siliceous algae (diatoms) is promoted. This explains, for example, the occurrence of thin beds of diatomaceous earth, alternating with tuff deposits, near Cheribon, south of Darma, Java, Dutch East Indies.

2. L. van Houten (1930) described radiolarite deposits, alternating with basic volcanic rocks, from the "Buchensteiner Schichten" (Ladinian) in the Dolomites. Next to these deposits large coral limestone reefs occur, indicating a shallow-water origin.

3. Dewey and Flett described Devonian radiolarite deposits from Cornwall, which alternate with lavas and shallow-water deposits.

4. Carstens described radiolarites from the Bymarck formation (Silurian) of the Trondhjemer syncline (Norway), which occur together with basic volcanic rocks of a shallow-water series.

5. J. B. Scrivenor (1912) described a Permo-Carboniferous section in Malaya which is called the "Pahang volcanic series." It is characterized by basic volcanic rocks (diabase tuffs, porphyries, porphyry tuffs, lahars, etc.) with clay slate intercalations and contains fossil plant remains and many radiolarites.

G. A. F. Molengraaff (1909) described similar formations from central Borneo, Dutch East Indies, as the "Danau formation." It is not so rich in plant remains but is very rich in radiolarites and manganese concretions (Molengraaff, 1915). The radiolarites alternate with coarse clastic sediments. J. Wanner (1921) and others, who believed in a deep-sea origin, supposed that these clastic sediments were deposited in small but very deep basins. But as shallow-water reef limestones also occur near by, this explanation seems erroneous.

The occurrence of many manganese concretions together with radiolarites is very characteristic of the Recent deep-sea radiolarian ooze, and this association is therefore often used as an argument for the deep-sea origin of sediments, for example, of the above-mentioned Danau formation of Borneo. However, the association of basic volcanic rocks and sedimentary manganese ore deposits is common in Turkey, for example. Such deposits were recently described by P. de Wijkerslooth (1943).

In northern Anatolia, in the so-called "Pontic geanticline" (Tromp, 1947, p. 368), most manganese ore deposits occur in the Upper Cretaceous flysch deposits, which are composed of an alternation of marls, andesites, and andesitic tuffs, in many cases rich in "jaspis" beds. According to Wijkerslooth, the original manganese deposition is due to volcanic exhalations. The manganese ore was later dissolved and redeposited in the form of concretions.

In the area between Ankara and Karabük

(central Anatolia), several small manganese deposits occur together with Jurassic radiolarites and serpentine massives, etc. The same association has been observed in the central Toros geanticline.

In the western part of the central Anatolian Plateau (Tromp, 1947, p. 368), the Upper Paleozoic beds are composed of limestones, graywackes, and radiolarites with many basic effusive rocks (diabase, etc.). Manganese interlayers are also quite common in these beds.

All these different phenomena sufficiently indicate that the association of volcanic rocks, radiolarian chert, and manganese concretions is a normal one, which might be due to deep-sea deposition but which, in general, is typical of shallow-water deposition connected with volcanic activity, either continental or submarine.

## REFERENCES CITED

HOUTEN, L. VAN (1930) Geologie des Pelmo Gebietes in den Dolomiten von Cadore, Thesis, Delft, Holland.

MOLENGRAAFF, G. A. F. (1909) Over oceanische Diepzeeafzettingen van Centraal Borneo, Akademie verslagen, Amsterdam, June 26.

——— (1915) Over Mangaanknollen in Mesozoische Diepzeeafzettingen van Borneo: Akademie verslagen, Amsterdam, vol. 23, pp. 1058–1073.

SCRIVENOR, J. B. (1912) Radiolaria bearing rocks in the East Indies: Geol. Mag., vol. 9, decade 5, pp. 241–248.

TROMP, S. W. (1947) A tentative classification of the main structural units of the Anatolian orogenic belt: Jour. Geology, vol. 55, pp. 362–377.

WANNER, J. (1921) Beiträge z. Geologie und Geographie von N. O. Borneo: Neues Jahrb., Beilage-Band 45, pp. 149–213.

WIJKERSLOOTH, P. DE (1943) Über die im weiteren Sinne sedimentären Manganerzlagerstätten West- und Zentralanatoliens: Maden Tetkik ve Arama Institüsü Bulletin, Ankara, sene 8, sayi 1/29, pp. 100–109.

# REVIEWS

*Outlines of the Geography, Life, and Customs of Newfoundland-Labrador (the Eastern Part of the Labrador Peninsula).* By V. TANNER. 2 vols. Cambridge: Cambridge University Press, 1947. Pp. 906; figs. 342.

In these two volumes Professor Tanner, of the University of Helsingfors, summarizes available information on the geography of eastern Labrador. Much original material, based on observations made during the Finland-Labrador Expedition in 1937 and the Tanner Labrador Expedition in 1939, is presented. Volume I is divided into five parts, which deal with the geology, oceanography, meteorology and climate, and plant and animal life; Volume II is given over entirely to human geography.

Geologists will be interested mainly in the first 254 pages describing the principal geologic features: the pre-Cambrian rocks and their structure, the Tertiary peneplain, effects of glaciation, and raised shorelines.

The summary of pre-Cambrian geology is based on studies by E. H. Kranck, a member of the 1937 expedition, and upon earlier studies by Daly, Coleman, Odell, and others. The following units are described in considerable detail:

Eruptive rocks of undetermined age
- Strawberry granite and related granites(?) and syenites

Algonkian series
- Double Mer and Lake Melville beds (Keweenawan)
- Ramah and Kaumajet (Cape Mugford) beds (Huronian)

Archean series
- "Domino gneiss"
- Makkobik granite
- Anorthosite gabbro
- Migmatite gneiss with mafic or lime-rich inclusions
- Migmatitic, microcline-rich gneisses and granite gneiss
- Aillik quartzite, conglomerate, and other sediments

Two major orogenic belts are recognized: (1) the Labrador mountain chain adjoining and paralleling the present coast, which was formed at the close of the Archaeozoic, and (2) the Labrador Range in the interior, which trends north-northwest from Mishikaman Lake to the west shore of Ungava Bay and was formed in late Huronian time.

The Cenozoic record, in which Tanner was particularly interested, begins with uplift of the "general peneplain surface" in the Pliocene. This surface is considered to be *metachronous* in the sense that it is composed of several facets developed at various times since the Huronian orogeny. The major preglacial valleys were initiated by uplift of the peneplain, and many persist to the present time. During the Pleistocene, Wisconsin ice from the west is believed to have covered the entire region, including the summits of the Torngak Mountains, which Daly, Coleman, and others previously had considered nunataks. No evidence of earlier glaciation is reported. Upwarp of the region following glaciation is evidenced by abundant raised shore features at various elevations. Because of uncertainties, only general interpretations are made from the marine limits, and there is no attempt to apply the epeirogenic spectrum graphs developed by the author in Fennoscandia.

As is clearly recognized, much of the geology of Labrador remains for future investigation. Professor Tanner's study, however, is a valuable summary and is of special significance because of the author's long experience with strikingly similar problems in Fennoscandia.

L. H.

---

*Aids to Geographical Research.* By JOHN KIRTLAND WRIGHT and the late ELIZABETH T. PLATT. ("American Geographical Society Research Series," No. 22.) 2d ed. New York: Columbia University Press, 1947. Pp. 331. $4.50.

The first edition of this useful reference work was published in 1923 as No. 10 of the "American Geographical Society Research Series" and has long been out of print. The new edition thus is completely revised and enlarged. It is primarily a guide to geographical bibliographies, periodicals, atlases, and gazetteers; but it also includes considerable geological reference mate-

rial, especially in the fields of physical geography and related sciences. The book is divided into three parts: Part I, "General Aids"; Part II, "Topical Aids"; and Part III, "Regional Aids and General Geographical Periodicals."

L. H.

---

*Correlation of Pleistocene Deposits of Nebraska.* By G. E. Condra, E. C. Reed, and E. D. Gordon. (Nebraska Geol. Survey Bull. 15 [1947].) Pp. 73; figs. 15.

Because of the unusually complete stratigraphic record of the Pleistocene found in Nebraska, this résumé is of general interest. The results of recent subsurface studies and soil investigations are emphasized, and a revised Pleistocene classification is presented. Two new stratigraphic names are proposed: (1) the *Seward formation*, as an eastern facies of the Ogallala and (2) the *Crete sand*, formerly described by Lugn as "valley phase" of the Loveland loess, as a channel deposit of Illinoian age. The loess deposits (Loveland, Peorian, and Bignell) are believed to be mainly eolian in origin, derived principally from silty Tertiary formations of the High Plains, and to be of interglacial rather than glacial age. Brief concluding sections deal with Pleistocene terraces, water-table fluctuations, economic relations, and plants and animals. Many Pleistocene geologists will not agree completely with the interpretations in the report, but most will find them stimulating and helpful.

L. H.

---

*1945 Reference Report on Certain Oil and Gas Fields of North Louisiana, South Arkansas, Mississippi and Alabama.* (*Shreveport Geol. Soc.*, vols. 1, 2.) Shreveport, 1947 (planographed). Vol. 1, pp. xxi+328; pls. 3. $12.50. Vol. 2, pp. xii+329–503, pls. 4–12. $10.00.

The greater portion of these two volumes is a summary, in telegram style, of the geologically important facts of 25 fields in Arkansas, 48 in Louisiana, 16 in Mississippi, and 1 in Alabama—a total of 90 fields. The data are grouped under the following subjects: location, prediscovery data, discovery well, well data, structure, producing zones, well spacing, allowable, development through January 1, 1946, deepest well in field, proven area, production data through January 1, 1946, pipe-line data, and miscellaneous. These data are tabulated on twelve pages following page 503, volume 2. Structure contour maps, cross sections, well logs, and other diagrams accompany the descriptions.

To students of Gulf Coast stratigraphy, two chapters by Hazzard, Blanpied, and Spooner (pp. 472–503 of vol. 2) on Cretaceous correlations and on the formations that underlie the Smackover limestone in southern Arkansas will be particularly interesting. New information from deep wells indicates that certain modifications of the earlier correlations will be necessary. Rocks of probable Permian age are believed to have been reached in several deep wells. Numerous correlation charts and diagrams illustrate the discussion.

The compilation of this immense amount of material has been difficult, as explained in the Preface, and several pages of errata had to be added. Interspersed with the text are advertisements of thirty-three firms. The typography of the books is good, but a few pages are missing, and some folded plates have been clipped in so that they must be cut to be used. The high cost of these volumes is most regrettable.

R. B.

---

*Papers from the Geological Department, Glasgow University.* (Glasgow University Publications, vols. 69–71; Geological Department Publications, vols. 20–22.) Glasgow: Jackson, Son & Co., 1947.

The first two volumes consist of twenty-seven reprints of octavo size originally published during the period 1937–1945. The third volume consists of twelve papers of larger page size, printed in 1937–1945 and now gathered into a quarto volume. The thirty-nine articles form a cross section (not a complete set) of the publications of the University's Geological Department and include a great range both of subject and of importance. Papers deal with such diverse branches of geology as paleontology and evolution (fifteen papers), structural geology (six papers), petrology (four papers), and smaller numbers concerned with geomorphology, economic geology, and techniques. They range in length and value from a page of strictly local interest to 65 pages with good discussions of principles, descriptions of new fossil species, or original rock analyses. Perhaps the greatest value of these volumes is that they bring to the attention of a larger circle of geologists some useful

papers originally printed in what are, to most American geologists at least, seldom seen publications.

M. S. C.

---

*Transactions of the Geological Society of South Africa*, vol. 47 (January to December, 1944), Johannesburg: Hortors Ltd., 1945. 2G.

This issue of the *Transactions* contains the following papers: (1) "Hillslopes and Dongas," by L. C. King and T. J. D. Fair; (2) "Fossils from the Pipe Sandstone at Victoria Falls, Rhodesia," by F. Dixey; (3) "The Geomorphology of Northern Rhodesia," by F. Dixey; (4) "The Bushveld Granites in the Zaaiplaats Tin Mining Area," by C. A. Strauss and F. C. Truter; (5) "The Basal Beds of the Transvaal System at Olifants River Poort, North East Transvaal," by J. W. Brandt and H. D. le Roex; (6) "Columnar, Conical, and Other Growths in the Dolomites of the Otavi System, S.W.A.," by C. M. Schwellnus and H. D. le Roex; (7) "Further Examples of Syntexis by Karroo Dolerite," by E. D. Mountain; (8) "The Origin of the Iron and Manganese Deposits in the Postmasburg and Thabazimbi Areas," by J. E. de Villiers; (9) "Stylolitic Solution in Witwatersrand Quartzites," Robert B. Young; (10) "Further Examples of Subterranean Subsidence of the Marievale Type and Syntexis Associated with Them," by J. Ellis; (11) "The Structure, Ore Genesis, and Mineral Sequence of the Cassiterite Deposits in the Zaaiplaats Tin Mine, Potgietersrust District, Transvaal," by P. G. Söhnge; (12) "The Vredefort Structure as Revealed by a Gravimetric Survey," B. D. Maree; (13) "Stratigraphic Features and Tectonics of Portions of Bechuanaland and Griqualand West," by D. G. L. Visser; (14) "Geomorphology of the Natal Drakensberg," by L. C. King; and (15) "A Comparison of the Gravimeter and Torsion Balance Methods of Geophysical Prospecting," by J. F. Enslin.

---

"On the Younger Pre-Cambrian Granite Plutons of the Cape Province." By D. L. Scholtz. (Anniversary Address by the President.) (*Geol. Soc. South Africa Proc.*, vol. 49.) Johannesburg, 1947. Pp. xxxv–lxxxii.

Dr. Scholtz's presentation covers the following topics: texture and mode of weathering of the granites, age of emplacement of the plutons, metamorphism of the Malmesbury sediments, inclusions in the granites, mineralogical composition of the granites and quartz porphyries, the structure of the plutons, differentiation and contamination of the granites, radium content of the granites, and the ore deposits. Use is made of Niggli's Q.L.M. diagram in the interpretation of the differentiation of the plutons, the conclusion being that gravitational differentiation, in conjunction with contamination, accounts for the increased basicity with depth. Evidence for a considerable amount of stoping is presented, and Wahl's ideas on thermal diffusion-convection differentiation are considered in accounting for the silica-alkali-alumina-rich border facies of the masses.

Fifty-two new chemical analyses are presented: forty-one of the pluton materials proper, five of contaminated and xenolithic material, and six of the various country rocks. Tables of recalculated chemical data, geologic and structural maps, photomicrographs and outcrop photographs, and variation diagrams are all presented.

In addition to the anniversary address, the 1946 volume of the *Transactions and Proceedings of the Geological Society of South Africa* contains the following papers: (1) "The Discovery and Prospecting of a Potential Gold Field near Odendaalsrust in the Orange Free State, Union of South Africa," by A. Frost, R. C. McIntyre, E. B. Papenfus, and O. Weiss; (2) "Notes on the Microscopic Features of the Magnetic Iron Ores of the Bushveld Complex," by C. A. Strauss; (3) "Corundum 'Indicator' Basic Rock and Associated Pegmatites in the Northern Transvaal," by J. W. Brandt; (4) "The Geology of a Portion of the Rooiberg Tinfields," by L. G. Boardman; (5) "Qualitative Spectrochemical Analysis of Minerals and Rocks," by L. H. Ahrens and W. R. Liebenberg; (6) "Petrofabric Analysis of the Bushveld Gabbro from Bon Accord," by J. J. van den Berg; (7) "Note on a Mechanical Method for the Quantitative Estimation of Graphite," by R. J. Adie; (8) "The Simple Dykes and Sills of the Far East Rand," by J. Ellis; (9) "The Development of the Vaal River and Its Deposits," by H. B. S. Cooke; (10) "Résumé of the Geology of the Richtersveld and the Eastern Sperrgebiet," by P. G. Söhnge and John de Villiers; and (11) "Die Korrelasie van Sekere Voor-Transvaal-Gesteentes in die Distrik Schweizer Reneke," by O. R. van Eeden.

Julian R. Goldsmith

## GLACIOLOGICAL WORK OF THE BRITISH JUNGFRAUJOCH RESEARCH PARTY

"A Crystallographic Investigation of Glacier Structure and the Mechanism of Glacier Flow." By M. F. PERUTZ and GERALD SELIGMAN. (*Royal Soc. London Proc.* ser. A, no. 950, vol. 172.) Pp. 335–360, 1939.

"The Temperature, Melt Water Movement and Density Increase in the Névé of an Alpine Glacier." By T. P. HUGHES and GERALD SELIGMAN. (*Royal Astron. Soc. Monthly Notice, Geophys. Supplement*, vol. 4.) Pp. 616–647, 1939.

"The Structure of a Temperate Glacier." By GERALD SELIGMAN. (*Geog. Jour.*, vol. 97.) Pp. 295–317, 1941.

"Mechanism of Glacier Flow." By M. F. PERUTZ. (*Phys. Soc. Proc.*, vol. 52.) Pp. 132–135, 1940.

"Growth of Glacier Crystals." By GERALD SELIGMAN. (*Nature*, vol. 161, no. 4091.) P. 485, March 27, 1948.

During the past decade and in spite of wartime interruption, British glaciological work has advanced rapidly, as attested by the birth of the *Journal of Glaciology* in January, 1947. Demorest's tragic death in 1942 was a severe setback to American work on the physics of glaciers, but current interest in ice and snow is mounting steadily in the United States and Canada. The value of a review dealing with data published nearly ten years ago and partly digested in Matthes' (1942*a*) excellent treatment on glaciers and in his earlier review of one of the papers (Matthes, 1942*b*) may not be apparent. However, it is felt that the full significance of the Jungfrau research work has been obscured by World War II and that a more complete recapitulation of the major findings may serve to kindle enthusiasm for further work in this promising field on this side of the Atlantic.

British contributions in glaciology during the last decade are largely the product of a research group organized and directed by Mr. Gerald Seligman which takes its name from the Jungfraujoch, an ice- and snow-covered saddle, at 3,460 meters altitude, in the névé region of the Great Aletsch Glacier of Switzerland. This site was selected for a research station because, in addition to its obvious natural fitness, easy access was afforded by the Jungfrau railway leading to a hotel on a near-by rocky peak (the Sphinx), which provided comfortable accommodations for the research party. A meteorological station atop the Sphinx constituted a further asset, and the sum total of these advantages unquestionably allowed this group to operate more effectively than could investigators working in a remote region where equipment is necessarily limited and a major share of the party's time and energy is devoted to the day-to-day problem of staying alive. A cavern excavated in the ice apron of the Sphinx and lighted by power brought from the railway served as a cold laboratory in which to study the physical properties and crystallography of ice and firn. Shafts and bore holes were sunk in near-by firn fields, principally the Mönchfirn at 3,460 meters altitude, and additional investigations were conducted in crevasses of the névé and in ice grottos of the glacier tongue. In none of these did the depth of penetration exceed 30 meters. Mean annual air temperature on the Mönchfirn is $-7°$ C.

For approximately 5 months between April and September, 1938, field investigations were carried forward on the following: (1) transformation of firn into glacier ice, (2) measurement of temperature distribution and variation in the névé area, (3) behavior and function of meltwater in firn, (4) measurements of firn density, (5) evaluation of the function of meltwater and compaction-settling in increasing firn density, (6) crystallographic studies of firn and glacier ice by petrographic methods, (7) mechanics of firn and glacier flow, (8) origin of ice bands in névé and of blue bands in glaciers.

Crystallographic studies by M. F. Perutz showed that most ice crystals in the uppermost layers of the firn were oriented with their *c*-axes nearly vertical, that is, parallel to the temperature gradient. This orientation developed from a dispersed arrangement in new-fallen snow by melting, recrystallization, and sublimation during the transformation from snow to firn because crystals oriented with their *c*-axes parallel to the temperature gradient grew at the expense of others not so oriented. It was also established that at a depth of 14 meters in the firn the vertical orientation had noticeably deteriorated and at a depth of 23 meters it was almost completely destroyed. At still greater depths the dispersed orientation was augmented. However, crystals in the ice bands of the firn maintained a predominant vertical orientation throughout the

depth of observation, and this was attributed to the immobility of the tightly grown crystals composing the bands. Loss of orientation in the loose firn was attributed to shifting of individual crystals or groups of crystals as compaction by settling occurred. This was confirmed by rectangular networks of pins embedded in the walls of shafts. These pins showed movements within the firn to be erratic in direction and amount and to be confined to individual crystals and groups of crystals. No evidence of movement along discrete slip planes was found, and lack of noticeable increase in grain size was thought to rule out recrystallization as a possible explanation for the development of a dispersed crystallographic orientation.

From thermocouples buried in bore holes and in the walls and floor of a shaft it was ascertained that firn temperatures are practically always zero below 15–20 meters, and the conclusion was reached that below these depths the entire glacier is at the pressure-melting temperature. Amelioration of the winter cold wave in the firn and changes of temperature therein were so rapid that they could not have been due wholly to conduction and were, therefore, attributed to percolation of meltwater. Thus, warming of the firn depended primarily on the supply of meltwater and the permeability of the firn as controlled by grain size, packing, density, impermeable crusts, and, above all, horizontal ice bands. By the end of summer, pressure-melting temperatures prevailed throughout the entire névé of the Aletsch Glacier.

Meltwater circulation in the firn was studied in some detail. Pans buried in firn near the surface collected more water than did those at greater depth unless local irregularities, such as ice bands, interfered. As expected, the amount of radiation falling on the surface proved to be the major factor controlling the supply of meltwater, but a fresh fall of snow also produced an increase, owing to the relatively rapid melting of the delicate snow crystals. During warm weather vertically descending meltwater at a depth of 1 meter amounted to 0.1–0.3 cc/cm²/hr. Flow of meltwater along firn layers dipping 8.5° was measured at 8.3 cm. per hour; but its amount, 0.005 cc/cm²/hr, was so small that the accuracy of the volume determination was questioned.

Density in the Mönchfirn at 3,455 meters altitude increased progressively from 0.3 in old snow at the surface to 0.6 in firn at 7 meters depth. Subsequent changes were much slower, the density reaching only 0.68 at 30 meters depth. In a crevasse at 3,300 meters altitude, the firn at 8 meters depth and age two years had a density of 0.63; at 14 meters and three years age the density was 0.72; and at 23 meters depth and 6 years age the density was 0.80. The density of firn is increased by freezing of meltwater and by compaction-settling accompanied by some sublimation and surface migration of water molecules. In general, early-summer freezing of descending meltwater which destroys the winter cold wave was considered more important than early-winter freezing of meltwater remaining in the firn from the preceding summer. The number of times a layer of firn is exposed to seasonal temperature changes before it becomes protected from such changes by deep burial is clearly significant. This is controlled by the annual rate of snow accumulation and depth of penetration of the winter cold wave. If accumulation is small and depth of penetration large, a firn layer will undergo many more seasons of freeze and thaw than if the reverse is true, and the increase of its density by freezing of meltwater will be correspondingly great. On the Mönchfirn a layer of firn undergoes only four or five seasons of freeze and thaw at the most.

Measurements at the beginning and end of the summer at a depth of 1.2 meters showed a density increase from 0.35 to 0.53. This was much greater than the increase accounted for solely by freezing of meltwater, and the difference was attributed to slow settling and compaction. The settling rate of new snow at a depth of 20 cm. was as much as 0.8 cm/cm/day when temperatures were above zero and about 0.13 cm/cm/day when temperatures were below zero. These rates of settling would produce an increase in density of new snow from 0.15 to 0.25 in 12 hours at temperatures above zero or in 3 days at temperatures below zero. The mean rate of settling of firn to a depth of 2.5 meters proved to be $1.9\text{–}2.25 \times 10^{-3}$ cm/cm/day. This is sufficient to produce an increase from 0.4 to 0.52 in 100 days. These observations provide the basis for concluding that settling and packing by shifting of crystals is the major cause of density increase in snow and firn; and the fact that no great increase in crystal size is noted, except in the initial stages, is cited as substantiating evidence that freezing of meltwater is a subordinate factor. Furthermore, calculations show that freezing of meltwater can account for only a small part of the total increase in density.

Two types of bands were recognized in the firn—the brownish annual bands and thin strata of clear ice of modest areal extent, one being traced over an area 150 × 160 meters. The annual bands are discolored by fine rock dust, spread over the glacier by wind in late summer, when bare rock is exposed on near-by mountain sides. In the course of excavating and maintaining shafts, it was discovered that the clear ice bands are formed within firn along relatively impermeable, high-capillarity crusts or layers which become saturated with downward-percolating meltwater and are frozen to solid ice by the next cold wave from the surface.

Firn is granular, compacted snow that is distinguished from glacier ice by intercommunicating air channels, which render it permeable to water. In glacier ice the intercommunicating air channels have been broken up and sealed off, so that the mass is relatively impervious. This difference is clearly demonstrated by Perutz's thin sections, which also show that air spaces in glacier ice are fewer, larger, and chiefly intracrystalline. The transition from firn to ice takes place with remarkable uniformity at a density between 0.82 and 0.84 and is not accompanied by a marked rise in density or crystal size.

Ice crystals in firn are rarely larger than 2 mm., but in the glacier tongue they are 1–10 cm. and occasionally more. The dispersed crystallographic arrangement in firn is also replaced by a preferred orientation in the glacier with the basal glide plane of many crystals lying parallel to the direction of flow. An increase in density to about 0.92 also occurs and is attributed largely to growth of crystals with elimination of air bubbles by plastic deformation during flowage. Crystal growth is attributed to the fact that crystals oriented with the basal glide plane parallel to the direction of shear have smaller internal stress and less free energy than do crystals of unfavorable orientation. The former therefore grow at the expense of the latter by transfer of molecules across crystal boundaries. Crystallographic studies showed no evidence of twinning or formation of new crystal nuclei during flow; and no trace of saline films was recorded, although the methods used were capable of detecting concentrations as low as 0.004 per cent NaCl and 0.008 per cent $NH_4NO_3$.

Lines of screws embedded in the ice wall of a grotto in the glacier tongue recorded actual slip along a blue ice band. Intracrystal air bubbles in ice adjoining the blue band were flattened, and the blue band itself consisted principally of large crystals from which air bubbles had been eliminated and concentrated in a thin layer just above the blue band. Most of these crystals were oriented with the basal glide plane parallel to the plane of the band. Blue bands were observed cutting sedimentary ice bands inherited from the névé, and the general conclusion was reached that they are of tectonic origin. From various crystallographic observations it was concluded that flowage in glacier ice is due to the plastic deformation and growth of individual crystals and to slip along fractures where the stress is sufficient.

Following wartime interruption, work is being continued by this group, with attention focused at present on the mechanism of crystal growth in glacier ice. It is felt that eventual solution of this problem will result in a fundamental contribution to the understanding of glacier flowage. Initial investigations show that crystals in flowing ice are not nearly so large as has been stated heretofore. Further publications from this ambitious glaciological program are awaited with anticipation.

## REFERENCES CITED

Matthes, F. E. (1942*a*) Glaciers, chapter v in Hydrology: physics of the earth, vol. 9, pp. 149–219, New York, McGraw-Hill Book Co., Inc.

——— (1942*b*) Review of The structure of a temperate glacier. By Gerald Seligman: Geog. Rev., vol. 32, pp. 342–344.

Robert P. Sharp

VOLUME 56 NUMBER 6

# THE JOURNAL OF GEOLOGY

*November 1948*

## CORRELATION OF PLEISTOCENE DEPOSITS OF THE CENTRAL GREAT PLAINS WITH THE GLACIAL SECTION[1]

JOHN C. FRYE, ADA SWINEFORD, AND A. BYRON LEONARD
State Geological Survey, University of Kansas, Lawrence

ABSTRACT

Integration of Pleistocene chronologies of the central Great Plains and the glaciated area is a major problem of late Cenozoic stratigraphy in North America. Lenticular deposits of volcanic ash associated with fossil mollusks occur in both regions and furnish a widespread datum for interregional correlations. The ash lentils, collectively called Pearlette, can be differentiated petrographically from other late Cenozoic ash deposits of the Plains region and have been studied at localities extending from southeastern South Dakota to northwestern Texas. The associated molluscan fauna possesses an unforeseen degree of uniformity and stratigraphic significance. The Pearlette ash and faunal zone occurs above Kansas till and below Loveland loess and Iowa till in the Missouri Valley region and is judged to be early Yarmouthian in age. A modification of stratigraphic names for Kansas contributes to uniformity of terminology in the Plains region.

### INTRODUCTION

The late Cenozoic glacial epoch has left some imprint of its history on virtually every region of the North American continent. The physical changes in the earth and the evolution of faunas during this time have been subjects of detailed study. In some regions, such as those covered by extensive continental ice sheets, the wealth of data recorded by many investigators has led to a detailed chronology and stratigraphic classification that is based primarily on advance and retreat of the various ice sheets.

Nevertheless, in any attempt to review the late Cenozoic history of the continent as a whole, one is immediately confronted with the lack of correlation among four distinct environmental provinces, each of which has its own chronology that is largely unrelated to that of other regions. These provinces are (1) the extensive area covered by the continental ice sheets, containing deposits made by ice and other deposits intercalated between them; (2) extra-glacial areas which were subjected to prolonged erosion during the epoch of successive glaciations, possessing a record that consists primarily of erosion surfaces rather than of sediments; (3) regions of fluvial and eolian deposits where the chronology has been developed largely (*a*) on the basis of fossil vertebrate faunas, as in the central and southern Great Plains, or (*b*) on the basis of an extensive terrace sequence, as in the lower Mississippi Valley region; and (4) areas of the continental margins containing marine deposits which have been differentiated largely on the basis of marine invertebrates.

It is the purpose of this paper to present data which establish a more precise

[1] Manuscript received April 23, 1948.

correlation of the glacial succession of the upper Mississippi Valley with fluvial and eolian strata of the central and southern Great Plains.

During the past decade and a half, detailed field studies and investigations of fossil vertebrate faunas have been made in Kansas and Nebraska. As a result of these studies in Kansas, large areas and considerable thicknesses of deposits formerly thought to be Pliocene in age and loosely classed as belonging to the Ogallala formation have been assigned definitely or tentatively to Pleistocene time. A review of the literature is not presented here but may be found for Iowa in Kay and Apfel (1928) and Kay and Graham (1943); for Nebraska in Lugn (1935), Condra, Reed, and Gordon (1947), and Schultz and Stout (1945; 1948, in press); for Texas in Evans and Meade (1945) and Meade (1945); and for Kansas in Frye (1945*a*, 1946). Although a usable scheme of chronology, based on stratigraphic and physiographic relations and fossil vertebrates, has evolved in the Plains region, the lack of adequate vertebrate faunas in deposits of the glaciated areas has prevented integration of this chronology with that of the glacial succession in the north-central United States, which has been accepted by many geologists as the standard Pleistocene record for North America. This lack of correlation with the glacial time scale has retarded clear understanding of Pleistocene chronology and stratigraphy in the Great Plains.

A step toward integration of the geologic record in the two provinces was taken when a buried soil zone that is part of the glacial series was traced from Iowa, South Dakota, and Nebraska southwestward into central Kansas (Lugn, 1935; Schultz and Stout, 1945, p. 241; 1948; Condra, Reed, and Gordon, 1947; Frye and Fent, 1947). This soil, which occurs at the top of the Loveland loess, establishes a key horizon of late Pleistocene age but still does not serve needs adequately because it is near the top of the succession of Pleistocene strata, and in some topographic situations it is subject to possible confusion with other buried soil zones.

A more important step toward integration of data representing the two provinces has been made possible by discovery of the widespread occurrence of distinctive deposits of volcanic ash within the late Cenozoic sediments of the Great Plains. In 1896 Cragin applied the name Pearlette to a volcanic ash lentil along Crooked Creek Valley in Meade County, Kansas. Similar deposits, some attaining a maximum thickness of more

---

PLATE 1

Pearlette volcanic ash bed in Iowa, Texas, and Kansas.

*A*, Exposure in gravel pit along the Missouri Valley bluff near Little Sioux, Iowa (loc. no. 3). Deposits classed by Iowa Geological Survey as Loveland sand and gravel at base (Grand Island member of Meade formation, Kansas classification), stratified sand, silt, and volcanic ash (Sappa member of Meade formation, Kansas classification), and Loveland loess with soil at top. Peorian loess occurs above the Loveland soil.

*B*, View at same locality as *A*, showing detail of Pearlette volcanic ash in stratified sand and silt and at top the supposed unconformity at base of Loveland loess.

*C*, Detail of slightly indurated Pearlette volcanic ash which caps a small mesa west of Channing, Texas (loc. no. 46).

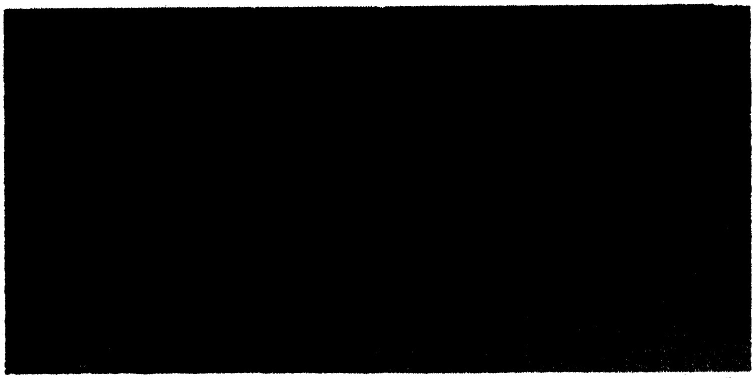
*D*, Meade formation in Gove County, Kansas (loc. no. 20). Base of Pearlette volcanic ash at hammer; beds in foreground yielded abundant snail fauna and consist predominantly of fragments of Niobrara chalk which immediately underlies the Mead formation at this locality.

A

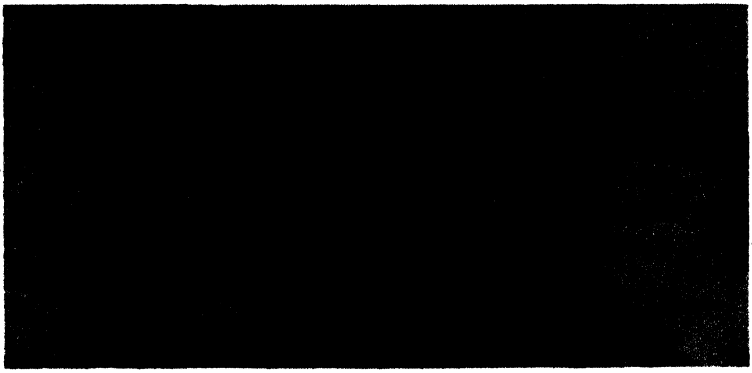
B

C

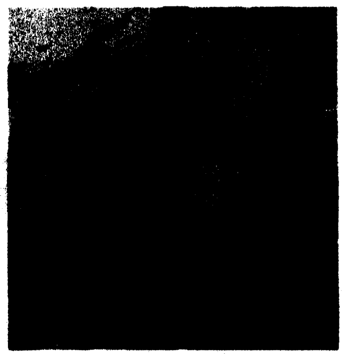
D



A



B



C D

than 20 feet, have been described from widely scattered localities in western Kansas (Landes, 1928), Nebraska (Lugn, 1935; Schultz and Stout, 1945, pp. 237–240), and Oklahoma, where volcanic ash has been mined commercially at many places (pl. 2). Most early workers in Kansas considered the ash to be Pliocene in age. Cragin's Pearlette lentil, however, occurs within the type area of the Meade formation, which has yielded several distinctive Pleistocene fossil vertebrate faunas (Hibbard, 1944).

Volcanic ash is also known to be interbedded with glacial deposits of the Missouri Valley region of South Dakota, Nebraska, and Iowa (Todd, 1892, 1895, 1899; Kay and Graham, 1944, p. 85; Schultz and Stout, 1945, p. 238; Condra, Reed, and Gordon, 1947, pp. 22–23). If there were means for distinguishing individual ash falls of different age, these lentils might serve as means of correlating deposits across the glacial border. The value of the ash lentils for correlation purposes is greatly increased by the association with them of both vertebrate and invertebrate fossil faunas.

In order to determine the distinctiveness of the several falls of ash, a petrographic and chemical study was made of volcanic ash deposits in western Kansas and adjacent states (Swineford and Frye, 1946) that had been approximately dated by vertebrate fossils and other methods. It was found that each of the several ash falls studied in the Great Plains possesses distinctive petrographic characters, and thus volcanic ash beds can be used for regional correlations.

During the summer of 1947 we examined in the field outcrops of the supposed Pleistocene volcanic ash throughout a belt extending from the Missouri Valley area of Iowa and Nebraska southwestward across Kansas and western Oklahoma to the described localities (Sidwell and Bronaugh, 1946) of northwestern and west-central Texas (pl. 1, *C*). Progress of field work was greatly facilitated by the active co-operation given to us by the state geological surveys of Iowa, South Dakota, Nebraska, and Oklahoma and the University of Nebraska State Museum. Dr. Raymond Sidwell, Texas Technological College, Lubbock, Texas, furnished us with samples of Pleistocene volcanic ash from localities in western Texas. Many of the localities in central and western Kansas had been studied prior to 1947, some of them in co-operation with Dr. C. W. Hibbard, formerly curator of vertebrate paleontology at the University of Kansas Museum of Natural History, and various members of the staffs of the state and federal geological surveys working in Kansas.

PLATE 2

Pearlette volcanic ash bed in Kansas, Oklahoma, and Nebraska.

*A*, Pearlette volcanic ash and thin overburden exposed in pit operated by Mid-Co Products Company south of Meade, Meade County, Kansas (loc. no. 36). This pit is in the vicinity of Cragin's type locality of the Pearlette ash. Snail fauna was collected from the clays below the ash and exposed in the floor of the pit.

*B*, Pearlette volcanic ash exposed in commercial pit north of Gate, Oklahoma (loc. no. 38). Snail fauna collected from clay exposed below the ash in near-by gullies. A pronounced north dip of the Pleistocene beds is due to solution-subsidence of the underlying Permian rocks.

*C*, Pearlette volcanic ash and underlying silty clay exposed in abandoned pit of the Cudahy Packing Company, near Orleans, Nebraska (loc. no. 9).

*D*, Same locality as *C*, showing Sappa (Upland), Loveland, and Peorian formations (Nebraska classification) exposed above the Pearlette volcanic ash.

Our field work was aided by two field conferences held during the summer of 1947. In May members of the state geological surveys and state universities of Iowa, Nebraska, and Kansas met in the field to study Pleistocene deposits exposed in western Iowa north of Council Bluffs, particularly at Loveland and near Little Sioux, and in the adjacent part of Nebraska. In June a regional field conference designed to study the stratigraphy of late Pleistocene deposits, particularly the loesses, convened at Peoria, Illinois, and worked westward across Illinois, Iowa, South Dakota, and Nebraska. This conference was attended by representatives of the state geological surveys of Illinois, Iowa, South Dakota, Nebraska, and Kansas, the University of Nebraska State Museum, the Soil Survey of the United States Department of Agriculture, and the Geological Survey of the United States Department of Interior.

Samples of ash were collected in the field from surface exposures by use of shovel, pick, and, in some cases, hand auger. The molluscan fauna, where present, typically occurs within the first few feet of sediments below the base of the ash (pl. 1, *D*) or, rarely, disseminated through the ash or immediately above it. Only those localities where the field relationships were clear have been included in this study. The sampling of the ash for an earlier study (Swineford and Frye, 1946) was quite thorough; many of the deposits were both channeled and spot sampled and the uniformity of petrographic and chemical characters demonstrated. Therefore, the sampling for this study consisted of the collection of one spot sample from a pure and unweathered portion of the ash lentil. All localities reported in table 1 and figure 1 were sampled by us except localities 4, 5, 21, 26, and 50. Locality 4 was sampled by A. G. Unklesbay of the Iowa Geological Survey, locality 5 by E. C. Reed of the Nebraska Geological Survey, locality 21 by Norman Plummer of the State Geological Survey of Kansas, locality 26 by A. C. Carpenter of Ottawa, Kansas, and locality 50 by Laura Lu Tolsted. In the field large sacks were filled with fossiliferous sediments; later, in the laboratory, the shells were separated from their matrix by washing the sediments through screens.

## PEARLETTE ASH ZONE IN THE GLACIAL SECTION

It is desirable, first, to establish the stratigraphic position of the Pearlette ash zone in the glacial succession of the Missouri Valley area. In 1924 Kay (p. 73) wrote concerning the well-known deposits of volcanic ash northwest of Little Sioux, Harrison County, Iowa: "The writer is convinced that this volcanic ash is not of Aftonian age, but is of the same age as the Loveland loess, with which in some of the county line exposures it is interstratified." In 1943 Kay and Graham (p. 85), in a paper on the Illinoian and post-Illinoian Pleistocene geology of Iowa, stated:

> The ash in the Little Sioux area is in the lower part of a loess phase of the Loveland formation, or may be associated with Loveland silts and clays showing deposition in water, but changing gradually into the typical eolian loess phase of the overlying Loveland. Frequently, sections of the Loveland which include the volcanic ash indicate that the pumicite was deposited in shallow ponds, lakes or bayous after the sands and gravels of the Loveland age had been deposited.

In 1947 Condra, Reed, and Gordon (p. 22) classed the Pearlette ash as a separate formation occurring conformably above the Upland[2] silty clay and the

[2] The Nebraska Geological Survey is now using the name Sappa to designate the deposits formerly

Grand Island sand and gravel (which constitute the lower part of the Loveland formation of Kay and Apfel and of Kay and Graham) and below the Crete sands and gravels and Loveland loess. During the field conference of May, 1947, the Pleistocene deposits exposed at the Loveland type section and in the vicinity of Little Sioux (pl. 1, *A* and *B*) were critically re-examined. It was agreed by the members of the conference that at the Little Sioux locality (1) thick stream-deposited sands and gravels occupy a deep erosional channel cut into glacial till of probable Kansan age; (2) these sands and gravels grade upward into stratified silts and sands that include the volcanic ash and associated snail fauna in its upper part; (3) laterally the channel gravels become thinner and the bedded sandy silts rest directly on glacial till; and (4) the bedded silts are overlain by two loess sheets, which are properly classed as Loveland and Peorian. The evidence in the Little Sioux area is less conclusive, however, as to the existence of an important unconformity at the top of the bedded silts that include the volcanic ash (Sappa) and the presence locally of thin channel deposits of sand and gravel (Crete) above the unconformity and below the Loveland loess.

The volcanic ash in Ringgold County, Iowa (loc. 4), is considered by members of the Iowa Geological Survey to occur above Kansas till, which in turn overlies Nebraska till, but relationships of the ash to the overlying beds are not clear.

Northward along the Missouri Valley, localities 1, 2, 5, and 6 are of considerable importance in establishing the placement of the ash zone, since at each of these localities either the ash lentils, the snail fauna, or both occur below a younger glacial till. At locality 2, in extreme northwestern Iowa, the volcanic ash was not found, but a large and distinctive snail fauna that is ascertained to be of contemporaneous age was collected from sediments stratigraphically above till sheets thought to be Nebraskan and Kansan in age and below Loveland loess, Iowa till, and Iowa loess. It should be noted that, although these localities all lie beyond the limits of the Iowa till as formerly mapped, they are well within the area shown by recent studies to have been covered by Iowan ice (Smith and Riecken, 1947; Flint, 1947; Warren, 1947; H. E. Simpson, 1947).

At locality 1 near Hartford, South Dakota, a thin lentil of volcanic ash occurs in a section of tills, loesses, and lenticular sand and gravel. The ash occurs above a small channel fill of sand and gravel resting unconformably on glacial till. Laterally, sand and gravel pinch out and the ash rests directly on weathered till. The ash is overlain by stratified silt and sand and Loveland loess; the latter displays a leached zone more than 1½ feet thick below unleached Iowa till. Iowa loess overlies the Iowa till. Dr. R. F. Flint, who is conducting a detailed investigation of the glacial deposits of South Dakota for the United States Geological Survey, was a member of the field conference when this section was examined and sampled.

At locality 5 in Knox County, Nebraska, the volcanic ash and associated snail fauna occur in stratified sand, silt, and clay (Sappa formation) that lie conformably on sand and gravel (Grand Island formation) which, in turn, rests on eroded Pierre shale. The beds containing the ash are overlain by Iowa till (per-

---

called Upland (personal communication from E. C. Reed, dated March 23, 1948), and throughout this paper the name Sappa will be applied to beds called Upland in earlier literature.

sonal communication from E. C. Reed). At locality 6, in Cedar County, Nebraska, volcanic ash occurs stratigraphically above Kansas(?) till and below Iowa till.

These data serve to establish the stratigraphic placement of the Pearlette volcanic ash and its associated molluscan fauna in the Missouri Valley region. The deposits containing the ash and fauna are shown to be younger than glacial till, judged to be of Kansan age, but older than Loveland lóess and overlying glacial till of Iowan age. Furthermore, prior to the deposition of the ash, the Kansas till was dissected by streams, and the valleys thus carved were deeply alluviated. Accordingly, the ash cannot belong to an age older than the Yarmouthian interglacial interval.

Placement of the ash with respect to the time of the Illinoian ice sheet, which advanced between Kansan and Iowan glaciations, is uncertain, largely owing to the lack of conclusive evidence concerning both the eastward correlation of the Missouri Valley Loveland loess and the presence of an erosional unconformity between the stratified beds containing the ash (Sappa) and the overlying Loveland loess. The problem of eastward correlation of the Loveland has been discussed by Kay and Graham (1943, pp. 47–49) as follows:

> Within the Illinoian drift area there are, in many places, two loesses on the Illinoian gumbotil and on surfaces of Illinoian drift. The younger of these two loesses is the Peorian loess; the older loess has been named in Illinois the late Sangamon loess. In Iowa this older loess has been correlated by Kay with the widespread Loveland loess of western Iowa, which is later than Kansan gumbotil erosion and is pre-Iowan in age. Leverett, however, correlated the Loveland loess of western Iowa with pre-Illinoian loess and questions the existence of a post-Illinoian, pre-Peorian loess. But in recent years this older loess on the Illinoian has been mapped widely by members of the Illinois and Iowa Geological Surveys. Although the Loveland loess of western, central, and southern Iowa outside the limits of the Illinoian area appears to be a single formation which was deposited in post-Illinoian, pre-Iowan time, Kay has stated that in reality its lower part may be pre-Illinoian in age, and only its upper part post-Illinoian; and it may be that a part of the Loveland loess in western Iowa and adjacent areas was deposited during the Illinoian glacial age.

In the Missouri Valley localities studied by us, a leached zone as much as 2 feet thick occurs in the top of the Loveland loess below unleached Iowa till, which suggests that even the upper part of the Loveland in this area cannot be considered younger than early Sangamonian. Furthermore, evidence from several localities (Condra, Reed, and Gordon, 1947, p. 24) indicates the presence of an unconformity between the Sappa formation of Nebraska and the Crete and Loveland formations of that area, even though this relationship is not clearly shown in the Little Sioux, Iowa, locality. We express our judgment that the weight of available evidence supports assignment of the Pearlette zone to the early part of the Yarmouthian interglacial age.

## METHODS OF CORRELATION

A detailed classification of Pleistocene time, based on the advance and retreat of the several ice sheets, has been in use for many years in the glaciated region. Lately, a usable but less detailed chronology has been developed in the Great Plains region, based primarily on fossil vertebrates, stratigraphic succession, and paleophysiography.

As already noted, relatively few vertebrate remains are found in the glaciated region, and thus vertebrate paleontological studies cannot be applied successfully to correlations of the glaciated and non-

glaciated areas. Stratigraphic succession, which is very useful in local areas, is not very helpful in regional studies, particularly where deposits of distinctive lithologic character, such as glacial till and buried soils, are lacking. Paleophysiography is a valuable tool in the study of fluviatile sediments, particularly in the Great Plains region, where major drainage changes have occurred, but, if work of this sort is to be wholly reliable, detailed knowledge of the entire drainage basin is required.

Two hitherto little-known tools for Pleistocene correlation have been used extensively in this study—the distinctive petrographic and chemical characteristics of volcanic ash beds and the composition of fossil snail faunas found to be widely distributed in association with ash deposits. Use of these tools will be described.

### PETROGRAPHY OF THE PEARLETTE VOLCANIC ASH

Samples of volcanic ash determined to represent the same fall as Cragin's Pearlette ash of southwestern Kansas were collected from forty-eight localities (fig. 1). The locations, occurrence, and petrographic data are given in table 1, and chemical analyses are given in table 2. The methods of petrographic study employed in the present investigation are essentially the same as have been described (Swineford and Frye, 1946) in previous work on Great Plains volcanic ash deposits. The laboratory procedures and distinctive characteristics of the ash are outlined briefly below. As the volcanic ash of the region is of the vitreous type, no mineralogical analyses are included.

*Laboratory procedure.*—Where several samples of volcanic ash were taken from one locality, the analytical results were reported for the sample which was the freshest and the least contaminated. The samples were dried and separated into the following size fractions for analysis: larger than 175 $\mu$, 175–125, 125–88, 88–62, and smaller than 62 $\mu$. Mechanical analyses were not made because previous determinations of particle-size distributions showed no significant pattern. In making the chemical analyses, portions of the 88–62 $\mu$ fraction were used, because this fraction in most samples was found to be relatively free from impurities and aggregates. The percentage of iron oxide, which was previously found to have diagnostic value, is reported in the petrographic table as well as in the table of chemical analyses. Megascopic color determinations of the portion passing the 62-$\mu$ screen were made by comparison with the Ridgway (1912) color plates.

The degree of alteration was studied (in the 175–88-$\mu$ fractions) by use of both the binocular and the petrographic microscopes. Particular note was made of the surface textures of the shards, the presence or absence of anisotropic boundaries which might interfere with refractive-index determination, the relative abundance of polarizing particles and inclusions, and the quantity of cemented aggregates resulting from decomposition.

Particle shape was studied (175–88 $\mu$) by use of a binocular microscope. Note was made of the relative thickness of the glass, the curvature of the bubble walls, the shape and number of junctures between bubbles, and the presence or absence of fibrous shards. Vesicles were noted under the petrographic microscope. Contaminants were studied under both microscopes and were reported for the 175–88-$\mu$ fractions. Refractive indices were determined by the immersion method with central illumination, using liquids at intervals of 0.002 and correcting for changes in temperature.

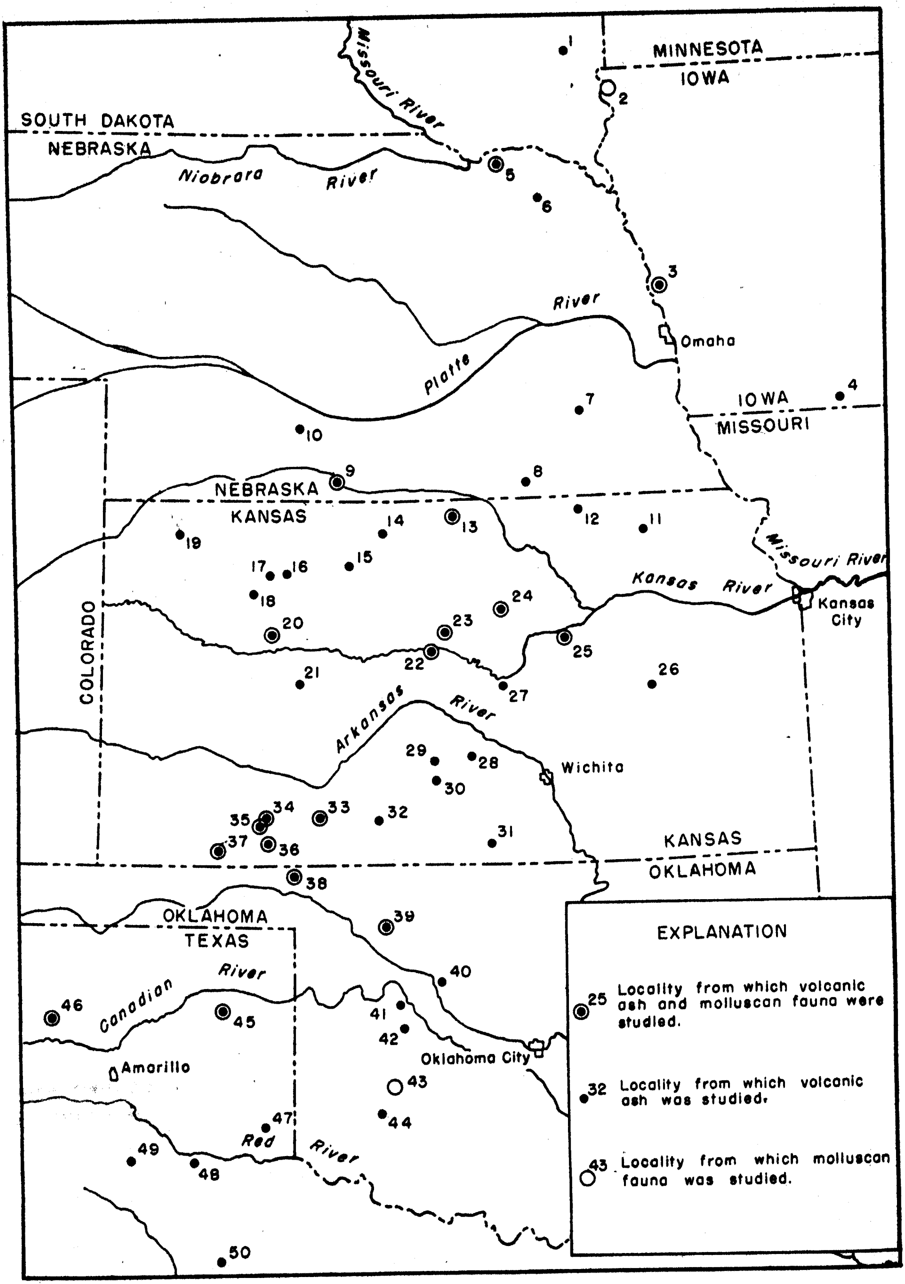



FIG. 1.—Map showing geographic distribution of the Pearlette volcanic ash and associated fossil snail localities studied. The numbered localities are listed in table 1.

TABLE 1

VOLCANIC ASH LOCALITIES STUDIED AND SIGNIFICANT PETROGRAPHIC CHARACTERISTICS OF THE VOLCANIC ASH SAMPLES

| No. | Location | Stratigraphic Position | Fauna | Color* ($<62\mu$) | Refractive Index | Per Cent $Fe_2O_3$ | Specific Gravity | Curved Junctures | Long Vesicles | Fibrous Shards | Alteration | Impurities |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | South Dakota | | | | | | | | |
| 2.... | Cen. NW. Sec. 13, T. 102 N., R. 51 W., Minnehaha Co. | 0.8 ft. ash, sample bottom 0.2; above Kan. till; below Lvld. loess & Iowa till | ............ | 17'''ff | 1.499–1.501 | 1.73 | 2.31 | c† | c | r† | Some crystalline inclusions | Tr. qtz. sd. |
| | | | | Iowa | | | | | | | | |
| 3.... | Sec. 6, T. 81 N., R. 44 W., Harrison Co. | 1.5 ft. ash, sample bottom 1 ft.; within waterlaid sd. & silt; above till; below Lvld. & Peorian loess | Snails below ash | 17'''ff | 1.499–1.501 | 1.96 | 2.28 | c | c | c | None | Tr. qtz. sd. |
| 4.... | SE. Sec. 18, T. 68 N., R. 28 W., Ringgold Co. | 1.5 ft. ash, road ditch, above brown clay & Kan. till; below loess | ............ | 17'''ff | 1.499–1.501 | 1.98 | ...... | c | c | r | Some crystalline inclusions; dull surfaces | Tr. qtz. sd. |
| | | | | Nebraska | | | | | | | | |
| 5.... | CN½ NW. Sec. 34, T. 33 N., R. 4 W., Knox Co. | Ash in stratified sd., silt, & clay; below Iowa till; above sd. & gr., above Pierre shale | Snails below ash | 19'''d | 1.499–1.501 | ...... | ...... | c | vc† | c | Many aggregates ($CaCO_3$); grain outlines somewhat anisotropic | 5% buff qtz. silt & sd. |
| 6.... | NE. Sec. 11, T. 29 N., R. 1 E., Cedar Co. | 2.5 ft. ash, in silt & sd.; above Kan. till; below Iowa till | ............ | 16'''ff | 1.499–1.501 | 1.68 | 2.33 | c | vc | vc | None | <1% qtz. sd. |
| 7.... | 400 ft. SW. of cen. Sec. 26, T. 9 N., R. 2 E., Seward Co. | 6 ft. ash, sample 3 ft. above base; in sdy., silty clay; above sd. & gr. | ............ | 17'''g | 1.499–1.501 | 1.81 | 2.23 | c | vc | c | None | Tr. qtz. sd. |
| 8.... | 2 mi. S., 0.1 mi. E., U.S. Hwy., Blue River bridge, Hebron | 2 ft. ash, road ditch; above gray, sdy. silt | ............ | 17'''f | 1.499–1.501 | 1.84 | ...... | c | r | vr† | Many crystalline inclusions; some anisotropic shard outlines; dull surfaces | Tr. qtz. sd. |
| 9.... | SE. Sec. 11, T. 2 N., R. 20 W., Harlan Co. | 8 ft. ash, sample 1 ft. above base; in sd. & silt; below Lvld. & Peorian loess | Snails above ash | 17'''g | 1.499–1.501 | 1.62 | 2.17 | c | c | c | Aggregates, 90% | 2–14% qtz. sd. |
| 10.... | S.½ NW. Sec. 20, T. 8 N., R. 24 W., Frontier Co. | >30 ft. ash; below Lvld. & Peorian loess | ............ | 17'''f | 1.498–1.501 | 2.08 | 2.25 | c | vc | c | None | Tr. qtz. sd. |
| | | | | Kansas | | | | | | | | |
| 11.... | NE. Sec. 11, T. 4 S., R. 9 E., Marshall Co. | 2.5 ft. ash, sample 2 ft. above base; creek bank, above gr., sd., & silt; below colluvium | ............ | 18'''f | 1.499–1.501 | 1.74 | ...... | vc | c | r | Aggregates, 95%; many crystalline inclusions; dull surfaces | Tr. qtz. sd. |
| 12.... | NW. cor. Sec. 20, T. 1 S., R. 3 E., Washington Co. | 1 ft. ash, above fn. sd. & silt; below red sd. | ............ | 17'''f | 1.499–1.501 | 1.61 | 2.34 | vc | c | vr | A few aggregates; some crystalline inclusions | None |
| 13.... | CN½ SW. Sec. 33, T. 1 S., R. 9 W., Jewell Co. | 3 ft. ash; above 3 ft. crs. to fn. sd.; above Niobrara chalk | Snails below ash | 17'''f | 1.499–1.501 | ...... | 2.27 | c | vc | c | None | Cretaceous foraminifers |
| 14.... | NW. Sec. 31, T. 3 S., R. 15 W., Smith Co. | >6 ft. ash, abd. creek meander | ............ | 17'''f | 1.499–1.501 | ...... | 2.29 | c | c | c | None | None |
| 15.... | SW. Sec. 18, T. 7 S., R. 18 W., Rooks Co. | 10 ft. ash, sample 5 ft. above base; above silt, sd., & gr.; above Niobrara chalk | ............ | 17'''g | 1.499–1.501 | 1.38 | 2.30 | c | vc | c | None | Tr. qtz. sd. |

* Color notations adapted from Ridgway (1912).

† vc, very common; c, common; r, rare; vr, very rare.

TABLE 1—*Continued*

| No. | Location | Stratigraphic Position | Fauna | Color* ($<62\mu$) | Refractive Index | Per Cent $Fe_2O_3$ | Specific Gravity | Curved Junctures | Long Vesicles | Fibrous Shards | Alteration | Impurities |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | Kansas—(*Continued*) | | | | | | | | | |
| 16 | Cen. W. side Sec. 11, T. 8 S., R. 25 W., Graham Co. | 9 ft. ash, abd. pit; above fn. sd. | ...... | 17'''f | 1.499–1.501 | ...... | 2.26 | c | vc | vc | A few crystalline inclusions | None |
| 17 | S. ½ Sec. 12, T. 8 S., R. 27 W., Sheridan Co. | 3 ft. ash, abd. pit, above sdy. silt, crs. gr., above Ogallala | ...... | 17'''ff | 1.499–1.501 | 1.93 | ...... | c | vc | c | Several aggregates; many crystalline inclusions | None |
| 18 | NW. Sec. 34, T. 8 S., R. 28 W., Sheridan Co. | 16 ft. ash, abd. pit; sample 3 ft. above base; above silt & sd., below massive silt | ...... | 17'''g | 1.499–1.501 | 1.84 | 2.26 | c | vc | c | None | None |
| 19 | SW. Sec. 14, T. 3 S., R. 35 W., Rawlins Co. | 10 ft. ash in gully, sample 3 ft. above base; above fn. sd., below silt | ...... | 17'''ff | 1.499–1.501 | 1.44 | 2.27 | c | c | c | None | None |
| 20 | NE. SW. Sec. 21, T. 13 S., R. 26 W., Gove Co. | 13 ft. ash, sample 5 ft. above base; above 6 ft. silt, sd., & gr.; above Niobrara chalk | Snails below ash | 17'''ff | 1.499–1.501 | 1.93 | 2.25 | c | c | c | None | None |
| 21 | S. ½ NW. Sec. 30, T. 18 S., R. 23 W., Ness Co. | >6 ft. ash overlain by silt | ...... | 17'''f | 1.499–1.501 | 1.47 | 2.32 | c | c | c | None | None |
| 22 | W. line SE. Sec. 35, T. 14 S., R. 11 W., Russell Co. | 0.7 ft. ash, road cut, sample from top 0.2; ash in stratified silt, sd., & gr. | Snails below ash | 17'''ff | 1.499–1.501 | 1.18 | 2.32 | vc | c | c | A few aggregates | Tr. qtz. sd. |
| 23 | Sec. 28, T. 13 S., R. 10 W., Lincoln Co. | 7 ft. ash, sample 2–5 ft. above base; in stratified silt, sd., & gr.; below silt | Snails below ash | 17'''g | 1.499–1.501 | 1.82 | 2.32 | c | c | c | None | Tr. qtz. sd. |
| 24 | NE. NW. Sec. 29, T. 10 S., R. 5 W., Ottawa Co. | 6 ft. ash, sample middle 1 ft.; above 3 ft. silt & clay; below 2 ft. silt, sd., & gr. | Snails below ash | 17'''g | 1.499–1.502 | 1.54 | ...... | c | r | r | Many crystalline inclusions; dull surfaces | None |
| 25 | SW. Sec. 35, T. 15 S., R 2 E., Dickinson Co. | 1 ft. ash; in silty clay; above & below sd., gr., & congl. | Snails above and below ash | 17'''f | 1.499–1.502 | 1.73 | 2.21 | c | vc | c | Crystalline inclusions, a few aggregates, and slightly dull surfaces | Tr. qtz. sd. |
| 26 | Cor. Sixth & West Sts., Emporia, Lyon Co. | 0.4 ft. ash, auger hole, 40 ft. below surface; below silt & clay & above sd. & gr. | ...... | 17'''f | 1.499–1.502 | ...... | ...... | c | c | c | Some crystalline inclusions; a few anisotropic boundaries; >50% aggregates | Tr. qtz. sd. |
| 27 | NW. SW. Sec. 15, T. 18 S., R. 4 W., McPherson Co. | 6 ft. ash, pit sample 2–4 ft. above base; in stratified sd. & silt; above sd. & gr.; below Lvld. silt | ...... | 17'''g | 1.499–1.501 | ...... | ...... | c | c | c | None | None |
| 28 | NW. cor. SW. Sec. 14, T. 25 S., R. 8 W., Reno Co. | 10 ft. ash, sample 3 ft. above base; above silt, sd., & peat; below silt, sd., & gr. | ...... | 17'''f | 1.499–1.501 | ...... | 2.31 | c | c | c | Some slightly anisotropic shard boundaries | None |
| 29 | NE. SW. Sec. 28, T. 25 S., R. 11 W., Stafford Co. | 2 ft. ash, auger hole in dune depression, sample 1 ft. above base; above gray clay, silt, & sd. | ...... | 17'''ff | 1.499–1.501 | 1.91 | 2.21 | vc | c | c | None | Tr. qtz. sd. |
| 30 | ½ mi. E. of SE. cor. Sec. 23, T. 27 S., R. 11 W., Pratt Co. | 2.5 ft. ash, sample 1.5 ft. above base; above silt, sd., & gr.; below silt | ...... | 17'''f | ? | ...... | ...... | c | ? | vc | Shards notably anisotropic; very dull surfaces | None |
| 31 | NE. cor. Sec. 29, T. 33 S., R. 6 W., Harper Co. | >6 ft. ash, abd. pit, sample 2 ft. above base; above silt & sdy. silt | ...... | 17'''f | 1.499–1.501 | 2.39 | ...... | r | c | c | Many crystalline inclusions; 95% aggregates | Qtz. sd., red silt |
| 32 | Cen. W. ½ Sec. 7, T. 31 S., R. 17 W., Comanche Co. | 8 ft. ash, sample 6 ft. above base; gray, sdy. silt & clay | ...... | 16'''ff | 1.499–1.502 | 1.57 | ...... | r | c | c | A few aggregates; shard surfaces slightly dull | Tr. qtz. sd. |
| 33 | Sec. 13, T. 30 S., R. 23 W., Clark Co. | 4 ft. ash, sample 2 ft. above base; above sdy. silt & sdy. gr.; below massive silt | Snails below ash | 17'''ff | 1.499–1.502 | 1.96 | 2.32 | c | c | vc | None | Tr. qtz. sd. |
| 34 | Sec. 2, T. 31 S., R. 28 W., Meade Co. | 17 ft. ash, pit sample, channel 7–9 ft. from base; above clay & sd.; below silt | Snails below ash | 17'''ff | 1.499–1.501 | 1.67 | 2.27 | c | c | c | Surfaces slightly dull | None |
| 35 | SE. SE. Sec. 33, T. 31 S., R. 28 W., Meade Co. | 8 ft. ash, sample 5 ft. above base; above sdy. silt | Snails below ash | 17'''ff | 1.499–1.501 | 0.06 | ...... | c | c | c | Crystalline inclusions; a few aggregates | Tr. qtz. sd. |
| 36 | NE. Sec. 26, T. 32 S., R. 28 W., Meade Co. | 18 ft. ash, pit sample 2 ft. above base; above clay & sd.; below silt | Snails below ash | 17'''g | 1.499–1.501 | 1.94 | ...... | c | c | vc | None | Tr. mica & qtz. sd.; a few brown insect parts |
| 37 | NW. NW. Sec. 35, T. 33 S., R. 32 W., Seward Co. | 7 ft. ash; above silt, sd., & gr.; below silt | Snails in ash | 17'''f | 1.499–1.501 | 2.07 | 2.31 | c | c | c | None | None |

| No. | Location | Stratigraphic Position | Fauna | Color* (<62μ) | Refractive Index | Per Cent $Fe_2O_3$ | Specific Gravity | Curved Junctures | Long Vesicles | Fibrous Shards | Alteration | Impurities |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | Oklahoma | | | | | | | | |
| 38.... | Sec. 8, T. 5 N., R. 28 E., Beaver Co. | 30 ft. ash; above silty clay; below silt | Snails below ash | 17''''g | 1.499–1.501 | 1.82 | ...... | ? | c | ? | None | ? |
| 39.... | NW. Sec. 23, T. 23 N., R. 18 W., Woodward Co. | 8 ft. ash, sample 2 ft. above base; above silt, sd., & gr.; below silt | Snails below ash | 17'''ff | 1.499–1.501 | 1.60 | 2.25 | c | c | c | Crystalline inclusions; some opal | 1% qtz. sd. |
| 40.... | E. ½ NW. Sec. 4, T. 18 N., R. 12 W., Blaine Co. | 6 ft. ash; sample 2 ft. above base; above silt, sd., & gr.; below sd., silt | ............ | 17'''ff | 1.499–1.501 | 1.77 | ...... | c | c | r | Shard outlines slightly anisotropic; surfaces dull | 10% qtz. sd. |
| 41.... | NE. Sec. 24, T. 16 N., R. 16 W., Dewey Co. | 4 ft. ash; above sdy. silt; below massive silt | ............ | 17'''f | 1.499–1.501 | 3.03 | 2.30 | vc | c | c | None | Tr. qtz. sd. |
| 42.... | SE. cor. NW. Sec. 15, T. 14 N., R. 16 W., Custer Co. | 15 ft. ash, pit sample 4 ft. above base; above clay, silt, & sd.; below 8 ft. red, stratified silt & gr. | ............ | 17'''ff | 1.499–1.501 | 1.34 | 2.30 | c | r | r | None | None |
| 44.... | SE. cor. Sec. 28, T. 6 N., R. 18 W., Kiowa Co. | 8 ft. ash; above silty clay; below 6 ft. massive silt & gr. | ............ | 17'''ff | 1.499–1.501 | 1.64 | 2.22 | c | c | c | A few crystalline inclusions | Tr. qtz. sd. |
| | | | | Texas | | | | | | | | |
| 45.... | NW. Sec. 14, Block B-1, Roberts Co. | 8 ft. ash, sample 4 ft. above base; above silt, clay, & sd.; below massive silt | Snails below ash | 16'''g | 1.499–1.501 | 2.07 | ...... | vc | c | c | A few crystalline inclusions | None |
| 46.... | 1 mi. N. of road, 8 mi. W. of Channing, Hartley Co. | >12 ft. ash capping butte, sample 3 ft. above base; above sds., silts, & clays | Snails below ash | 16'''g | 1.500? | 2.64 | 2.26 | c | vc | vc | Shard outlines anisotropic; surfaces dull | Tr. qtz. sd. |
| 47.... | SE. Sec. 90, Block 14, Collingsworth Co. | 25 ft. ash, sample 4 ft. above base; above fn. sd. & silt | ............ | 17'''f | 1.499–1.501 | 1.58 | 2.28 | c | r | c | Slightly anisotropic outlines | Tr. qtz. sd. |
| 48.... | Edge caprock on Clarendon-Silverton Hwy., 13 mi. SW. of Antelope Flat, Briscoe Co. | 12 ft. ash, sample 8 ft. above base; above 10 ft. sd. & silt, above Ogallala fm.; below massive silt & sd. | ............ | 17'''f | 1.499–1.501 | 1.69 | 2.32 | c | c | c | None | None |
| 49.... | 11 mi. E. of Tulia, Swisher Co. | 4 ft. ash, sample 1 ft. above base; above sd. & silt; below sd. | ............ | 17'''g | 1.499–1.501 | 1.65 | 2.28 | c | vc | c | Crystalline inclusions; a few aggregates | Tr. qtz. sd. |
| 50.... | 2 mi. N. of McAdoo, Dickens Co. | 10 ft. ash; above silty clay & silty sd. | ............ | 16'''ff | 1.499–1.501 | 0.80 | 2.31 | c | c | c | None | Tr. qtz. sd. |
| | | | | Pliocene Ash | | | | | | | | |
| *A*.... | NW. SW. Sec. 23, T. 2 S., R. 22 W., Norton Co., Kan. | Pliocene [Ogallala fm.] (Swineford & Frye, 1946, p. 16) | ............ | NGf | 1.500–1.502 | 2.16–2.42 | 2.33–2.37 | r | vr | vr | None | Tr. qtz. sd. |
| *B*.... | N. line NW. NE. Sec. 30, T. 1 S., R. 19 W., Phillips Co., Kan. | Pliocene [Ogallala fm.] | ............ | 17'''ff | 1.503–1.505 | 1.86 | 2.35 | c | vr | vr | Crystalline inclusions | None |
| *C*.... | 100 yds. N. of S. line SE. SE. Sec. 8, T. 14 S., R. 38 W., Wallace Co., Kan. | Pliocene [Ogallala fm.] | ............ | 10'''ff | 1.503–1.505 | 2.17 | 2.35 | c | vr | vr | None | None |
| *D*.... | Hemphill Co., Tex. (Reed & Longnecker loc. 20) | Pliocene [Ogallala fm.] (Swineford & Frye, 1946, p. 17) | Hemphill fauna below ash | NGg | 1.495–1.497 | ...... | ...... | vr | vr | vr | Many crystalline inclusions; very dull surfaces; many aggregates | Qtz. sd. |
| | | | | Miocene Ash | | | | | | | | |
| *E*.... | South-central Sioux Co., Neb., *Merychippus* Draw | Miocene [Sheep Creek fm. type locality] | ............ | 16''''dd | 1.505–1.507 | 7.22 | 2.32 | vr | vr | vr | Very slight | Tr. qtz. sd. |

Specific gravity determinations of thirty-two samples containing little detrital material or alteration products were made by pycnometer. The size grade 125–88 $\mu$ was used.

The results are reported in table 1. Five samples of older ash are included at the end of the table for comparative purposes. In comparing the results of chemical analyses and color determinations, more weight should be given to those samples which are fresh than to those in

TABLE 2

CHEMICAL ANALYSES OF SAMPLES OF VOLCANIC ASH*

| No.† | $SiO_2$ | $Al_2O_3$ | $Fe_2O_3$ | CaO | MgO | Ignition Loss‡ | Undetermined Difference | Moisture‡ |
|---|---|---|---|---|---|---|---|---|
| 1 | 71.41 | 14.00 | 1.73 | 0.78 | 0.06 | 4.12 | 7.90 | 0.30 |
| 3 | 70.98 | 13.01 | 1.96 | 1.16 | 0.22 | 3.91 | 8.76 | 0.29 |
| 4 | 69.88 | 15.25 | 1.98 | 0.52 | 0.37 | 4.83 | 7.17 | 0.59 |
| 6 | 67.87 | 13.04§ | 1.68 | 0.66 | 5.58 | 4.12 | 7.05 | 0.41 |
| 7 | 71.29 | 13.29 | 1.81 | 0.82 | 0.21 | 4.06 | 8.52 | 0.30 |
| 8 | 70.44 | 13.07 | 1.84 | 0.87 | 0.26 | 4.61 | 8.91 | 0.31 |
| 9 | 75.02‖ | 11.17 | 1.63 | 3.72 | 0.40 | 4.40 | 3.67 | 0.95 |
| 10 | 70.46 | 13.48 | 2.08 | 1.31 | 0.34 | 3.94 | 8.19 | ........ |
| 11 | 69.42 | 14.64 | 1.74 | 2.38 | 0.18 | 4.52 | 7.12 | 1.00 |
| 12 | 71.84 | 14.64 | 1.61 | 0.96 | 0.13 | 3.87 | 6.95 | 0.90 |
| 15 | 71.68 | 12.88 | 1.38 | 0.88 | 0.56 | 3.73 | 8.89 | 0.41 |
| 17 | 71.65 | 12.33 | 1.93 | 1.30 | 0.81 | 4.40 | 7.58 | 0.10 |
| 18¶ | 72.83 | 14.38 | 1.84 | 0.56 | ........ | 5.53 | 4.86 | ........ |
| 19 | 70.85 | 13.34 | 1.44 | 0.84 | 0.13 | 3.46 | 9.94 | 0.31 |
| 20 | 71.68 | 13.20 | 1.93 | 0.93 | 0.13 | 4.21 | 7.92 | 0.21 |
| 21¶ | 71.16 | 11.70 | 1.47 | 1.46 | 0.38 | 3.86 | 9.97 | ........ |
| 22 | 71.99 | 13.51 | 1.18 | 1.05 | 0.13 | 3.90 | 8.24 | 0.32 |
| 23¶ | 72.77 | 13.87 | 1.82 | 0.87 | ........ | 3.15 | 7.52 | ........ |
| 24¶ | 73.30 | 14.46 | 1.54 | 1.00 | 0.21 | 4.60** | 5.64†† | ........ |
| 25 | 69.92 | 13.62 | 1.73 | 1.13 | 0.71 | 4.55 | 8.34 | ........ |
| 29 | 71.88 | 14.12 | 1.91 | 0.95 | 0.33 | 4.22 | 6.39 | ........ |
| 31 | 66.27 | 16.09 | 2.39 | 1.26 | 0.06 | 5.22 | 8.71 | 1.68 |
| 32 | 72.16 | 12.94 | 1.57 | 0.84 | 0.20 | 3.79 | 8.50 | 0.74 |
| 33 | 70.40 | 12.66 | 1.96 | 0.99 | 0.02 | 4.08 | 9.89 | 0.34 |
| 34¶ | 72.64 | 12.06 | 1.67 | 0.74 | ........ | 4.24 | 8.65 | ........ |
| 35 | 75.83 | 13.69 | 0.06 | 1.60 | 0.16 | 4.62 | 4.04 | 0.55 |
| 36 | 71.98 | 13.52 | 1.94 | 0.62 | 0.30 | 3.55 | 8.09 | 0.17 |
| 37¶ | 74.34 | 12.40 | 2.07 | 0.64 | 0.65 | 4.24 | 5.59†† | ........ |
| 39 | 69.97 | 14.15 | 1.60 | 0.85 | 0.27 | 4.47 | 8.69 | ........ |
| 41 | 71.31 | 17.31 | 3.03 | 1.01 | ........ | 4.00 | 3.34 | ........ |
| 42 | 72.53 | 14.14 | 1.34 | 1.19 | 0.37 | 4.20 | 6.23 | ........ |
| 44 | 67.72 | 14.16 | 1.64 | 0.95 | 0.86 | 4.66 | 10.00‡‡ | ........ |
| 45 | 71.74 | 14.22 | 2.07 | 1.11 | 0.43 | 4.16 | 6.27 | 0.34 |
| 46 | 72.16 | 16.96 | 2.64 | 0.51 | ........ | 5.01 | 2.72 | 0.94 |
| 47 | 66.83 | 14.95 | 1.58 | 1.07 | 0.45 | 4.59 | 10.53§§ | ........ |
| 48 | 70.80 | 12.40 | 1.69 | 0.56 | 0.32 | 4.16 | 10.07 | 0.09 |
| 49 | 69.24 | 12.66 | 1.65 | 1.36 | 1.96 | 4.69 | 8.44 | 0.28 |
| 50 | 70.70 | 12.35 | 0.80 | 1.17 | 2.83 | 3.75 | 8.40 | 0.15 |
| B | 69.77 | 12.88 | 1.86 | 1.41 | 1.02 | 5.26 | 7.80 | ........ |
| C | 68.89 | 12.94 | 2.17 | 1.04 | 0.64 | 4.96 | 9.36 | ........ |
| E | 59.32 | 24.93 | 7.22 | 2.52 | 0.95 | 5.08 | −0.02 | ........ |

* Analyses by Russell Runnels in the laboratories of the State Geological Survey of Kansas.
† Sample numbers refer to localities shown in fig. 1 which are listed in table 1.
‡ Samples dried to 140° C. and ignited to 1,000° C.
§ Includes 0.87 per cent $TiO_2$.
‖ Contains quartz sand.
¶ Analysis by Frances Schloesser (Swineford and Frye, 1946).
** Includes moisture.
†† Alkalies determined
‡‡ Includes $K_2O$, 5.07 per cent; $Na_2O$, 5.09 per cent.
§§ Includes $K_2O$, 5.00 per cent; $Na_2O$, 5.25 per cent.

which alteration has taken place or to those which have been contaminated by sand or silt.

*Distinctive features of Pearlette ash.*—In a previous comparison of Pearlette ash with various ash deposits of Pliocene age in Kansas, the following properties were found to be diagnostic of the Pearlette:

1. The color includes certain light shades of orange-yellow, which are not characteristic of fresh ash described from the Pliocene (Calvert mine).

2. The refractive index is consistently 1.499–1.501.

3. The shape of the shards is characteristically sharply curved, with thickened glass at the bubble junctures, which are commonly curved and branching at wide angles. Fibrous shards are present in all samples.

4. Many shards have groups or clusters of elongate vesicles which are seldom found in ash of Pliocene age.

5. The percentage of iron oxide is less than 2 as shown by eight of nine chemical analyses, with a range from 1.43 to 2.07 per cent; whereas in fifteen analyses of samples of Pliocene ash the $Fe_2O_3$ content ranges from 1.66 to 3.09, with only two samples below 2.00 per cent.

6. The specific gravity ranges from 2.21 to 2.32, whereas in the Pliocene ash it ranges from 2.33 to 2.37.

The color of most of the Pleistocene ash samples examined for this study is characteristic of the Pearlette, although in some it is modified by alteration of the glass. The refractive index corresponds to that of the type Pearlette in forty-six of the analyses; two samples were altered in such a way as to make determination doubtful.

The percentage of ferric oxide ranges from 0.06 (which is unexplained) to 3.03, six of the thirty-five samples analyzed showing an iron oxide content of more than 2.00 per cent. The high value for one of them (no. 34, 2.39 per cent) probably is due to the presence of red silt as an impurity. Sample 46, which has 2.64 per cent $Fe_2O_3$, may have gained in percentage of iron through alteration. The values for three other samples (nos. 10, 37, and 45, 2.08, 2.07, and 2.07 per cent, respectively) are only slightly higher than 2.00 per cent. No. 41, however, which shows a value of 3.03 per cent, is anomalous. It is thought to be Pearlette because its other parameters are the same as those of typical Pearlette.

The specific gravity ranges from 2.17 (altered sample 9) to 2.34, only one being higher than the range previously described for Pearlette ash. This sample, no. 12, also has fewer fibrous shards than typical Pearlette, but it is tentatively classed with that ash fall. The vesicles and the shapes of the shards of all samples (with the possible exception of no. 12) are of the Pearlette type.

The chemical analyses of Pleistocene ash in table 2 are very similar, except for magnesium oxide. The percentage of MgO is less than 1.00 except for three samples (6, 49, and 50), which consist of relatively fresh, unaltered ash; the large quantities of MgO are not explained.

In the present study, samples previously described from the Ogallala have been included, and samples from two more ash falls have been obtained in order to make further comparisons. They are described at the end of table 1.

*A* is Pliocene Ogallala ash from the Calvert mine, Norton County, Kansas. *B* and *C* represent another ash fall within the Pliocene Ogallala formation exposed in Phillips and Wallace counties, Kansas. *D* is another Ogallala ash from Hemphill County, Texas. *E* is ash of late Miocene age from the type locality of the Sheep Creek formation at *Merychippus* Draw in

Sioux County, Nebraska; it was submitted for our examination by C. Bertrand Schultz and T. M. Stout of the University of Nebraska State Museum. All differ in at least two respects from the Pearlette and from each other. Although the refractive index of *A* is close to that of the Pearlette, the color is a distinct gray, the iron content is high, and the shards are flatter than Pearlette shards and they lack elongate vesicles. *B* and *C* have the color and shape of Pearlette ash particles, except for the absence of fibrous shards, but have a higher refractive index and few elongate vesicles. *D*, a weathered sample, can be differentiated definitely from the Pearlette by its lower refractive index and its flat platy shards. *E* has almost nothing in common with the Pearlette; its refractive index is much higher, its color is a dark gray, most of its shards are characterized by two or more parallel, nearly straight bubble junctures, and vesicles are rare; its chemical analysis shows a low percentage of silica and a large amount of iron.

In using physical and optical properties of volcanic glass for correlation, it is necessary to assume that the characteristics of the glass from one ash fall (the Pearlette) will differ perceptibly from those of the glass in other ash falls in the same period and from the same source. Such properties as the refractive index, specific gravity, and shape of shards depend not only upon the chemical composition of the magma but also upon such highly variable factors as the temperature of quenching, pressure, and gas content. Glass from acid magmas in particular may show much variation because of a wide range in temperature and other conditions at eruption (E. F. Osborn, oral communication). Therefore, it is unlikely that the glass shards from several different ash falls will have the same characteristics.

All other glasses studied in the central Great Plains which have a similar chemical composition to that of the Pearlette and are known to have been deposited from different ash falls can be distinguished from each other and from the Pearlette ash (table 1).

### MOLLUSCAN FAUNA

*Association of fauna with ash.*—Assemblages of fossil mollusks occur in association with Pearlette volcanic ash at localities (fig. 1) distributed from Nebraska (loc. 6) and western Iowa (loc. 3) to northwestern Texas (loc. 46). Fossil mollusks were studied from eighteen localities where the snails were in immediate association with the ash and from a similar stratigraphic position at two localities (locs. 3 and 43) where ash did not occur. The molluscan fauna from localities 3 and 43 was later shown to possess the distinctive features of the Pearlette faunal assemblage.

The molluscan fauna of the Pearlette zone was first studied in southwestern and central Kansas (Frye, Leonard, and Hibbard, 1943) where fossils are abundant in blue-gray to tan silty clays, occurring from a few inches to 2 feet below the base of the volcanic ash lentils. In these areas, fossil remains of small mammals and other vertebrates occupy the same zone; they have been listed and described by Hibbard (1944), particularly from Meade, Russell, and Lincoln counties, Kansas. At other places (locs. 13 and 20) the snails are dispersed through as much as 5 feet of silt and sand below the ash; rarely, as in Seward County, Kansas (loc. 37), the mollusks are deposited with the ash itself. In the old Cudahy ash mine, near Orleans, Nebraska (loc. 9), the fossils were collected from near the top and just above the main bed of ash but below thin lenses of ash in the overlying silts of the Sappa

formation. At locality 25, in Dickinson County, Kansas, the molluscan fauna occurs both below and above the ash stratum in tan to gray silty clay with which the ash is interstratified.

Although the mollusks associated with the Pearlette volcanic ash consist of both aquatic and terrestrial forms, seemingly without exception they were deposited in depressions or shallow ponds; the aquatic forms were, no doubt, residents of the ponds themselves, whereas the terrestrial forms probably were washed in from near-by land areas. In some places (loc. 3) the ponds or lakes were so large that offshore deposits contain only aquatic species; the faunule of shoreline deposits consists predominantly of terrestrial gastropods. At most places the small size of ponds allowed a mixture of the forms from the two environments, although a few deposits contain a faunule comprised entirely of aquatic species or entirely of terrestrial forms.

*Distinctive features of the fauna.*—The occurrence and distribution of the sixty-five species and subspecies which have been distinguished in the Pearlette molluscan fauna are shown in figure 2, where species of the faunules are divided into five categories: (1) species ranging from lower Pliocene to Recent, (2) species ranging from Blancan to Recent, (3) species ranging from Blancan into the Pearlette faunal zone but not higher, (4) species restricted to the Pearlette zone, and (5) living species which make their first known appearance in Yarmouthian deposits. "Living species" is interpreted to mean forms which are living at some place within North America; many of these no longer live in the area under consideration.

No review of the literature concerned with Pleistocene fossil mollusks is given here, inasmuch as the Pearlette fauna will be reported upon in detail in a study now in preparation. All specimens which comprise the basis for this report and the faunal paper to follow are catalogued in the molluscan collections of the University of Kansas Museum of Natural History.

It is obvious that assemblages of any fossil organisms are useful for purposes of stratigraphic correlation only if an assemblage, associated with some specific stratigraphic unit, possesses at the same time features which distinguish it from assemblages of similar organisms that are known to occur in deposits of different stratigraphic placement. Furthermore, the usefulness of distinctive elements of an assemblage is limited by their regional distribution, unless it can be shown that certain species of one region are comparable in evolutionary development to those in another region at the same stratigraphic level or unless it can be shown that the changes in elemental composition of an assemblage, from one region to another, are due entirely to regional variation in climate or to other environmental factors.

In order to discriminate between distinctive and less meaningful elements of the Pearlette molluscan fauna and to assist in evaluating the various components of this fauna, the species have been grouped under the five categories previously mentioned. This grouping is intended to emphasize two features of the fauna: (1) the vertical range of species and (2) the regional distribution of species, especially with regard to occurrence in the glaciated portion of the region under study and the occurrence in the nonglaciated Plains portion of the region.

The four species (fig. 2, nos. 1–4) which range from the early Pliocene Laverne formation (Leonard and Franzen, 1944) to the Recent are of little significance for present purposes, except that nos. 2, 3, and 4 are distributed across

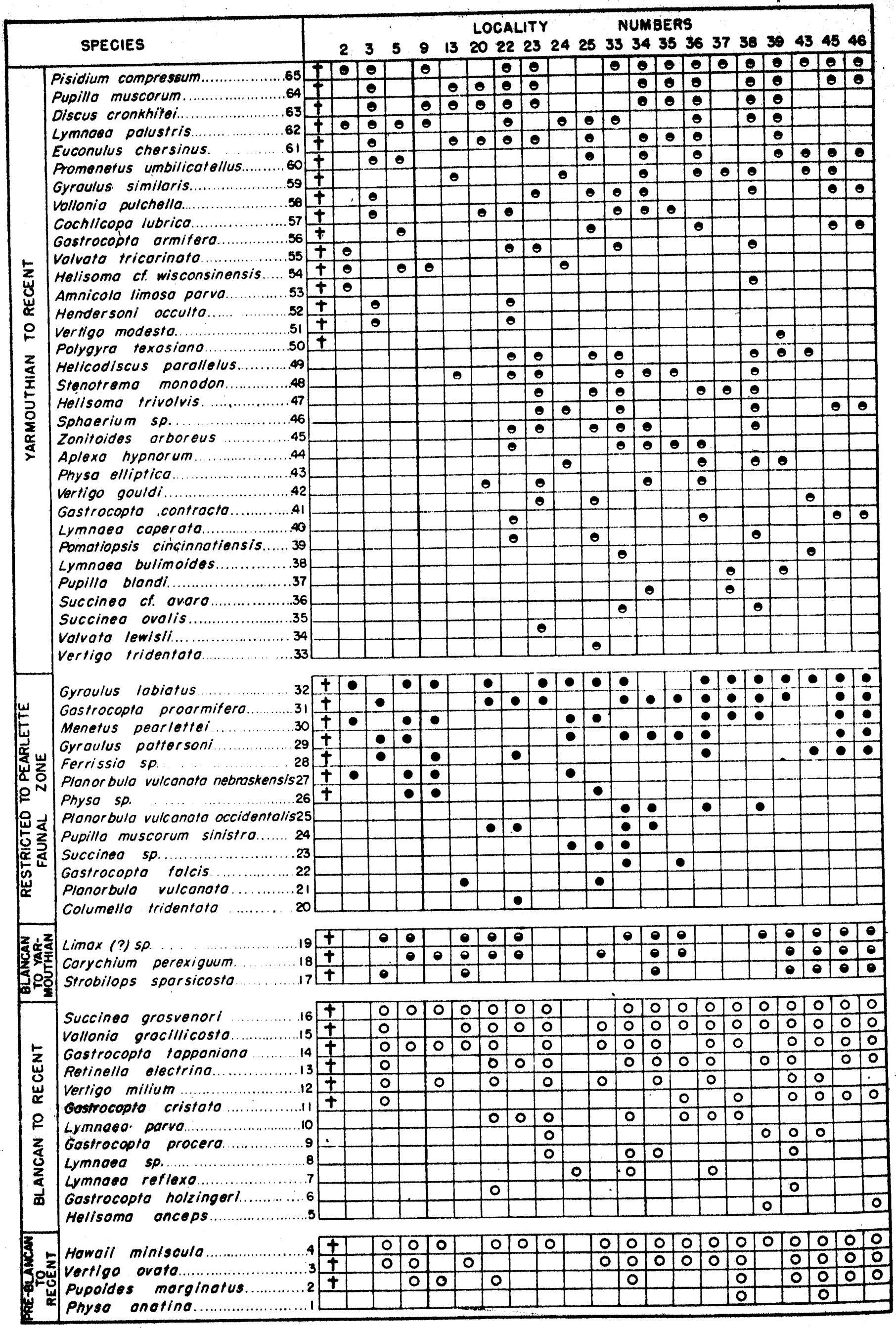



FIG. 2.—Chart showing distribution and occurrence of the 65 species comprising the Pearlette faunal assemblage. The species are grouped according to their vertical range. Species marked † are those which occur on either side of the glacial border.

the glacial border and serve to indicate widespread uniformity of ecological conditions at the time the beds containing them were being deposited. The twelve species in the next category (fig. 2, nos. 5–16) are of similar significance; the local occurrence of members of both groups does, however, contribute to a knowledge of environmental conditions at a specific place.

Three species (fig. 2, nos. 17–19) are not known to occur in deposits younger than those associated with the Pearlette ash. However, since they are known from Blancan deposits, they are useful as indices only when found directly associated with the Pearlette ash or when by other means it can be shown that the deposits are not Blancan in age. The fossilized internal shells of a large slug (no. 19), belonging to the molluscan family Limacidae, are of particular interest, since there are no living representatives of this family in the United States which possess internal shells of the type found as fossils in the Pearlette fauna. These shells have not been found in deposits of undoubted Pliocene age, but they are common in Blancan deposits.

Thirteen species or subspecies (fig. 2, nos. 20–32) are, as far as known, restricted in their occurrence to the Pearlette zone. This group is of great significance for purposes of regional correlation of the deposits containing them, since their vertical distribution is limited, and their value is further enhanced by the fact that a majority of this group (nos. 26–32) occurs both in localities within the glaciated portion of the region under study and in localities in the Plains region.

The largest group of species (fig. 2, nos. 33–65) is composed of those which range upward to Recent deposits (Leonard and Frye, 1943; Hibbard, Frye, and Leonard, 1944) but which make their first known appearance in the Pearlette faunal zone. Among these fossils are representatives of seven genera (*Cochlicopa*, *Discus*, *Pupilla*, *Stenotrema*, *Vallonia*, and *Valvata*) which are not known to occur in older faunal zones within the area under study. Sixteen of the species in this group (fig. 2, nos. 50–65) have wide distribution and occur in localities on either side of the glacial border. Since these species range upward into younger deposits, they are useful for stratigraphic correlation only when definitely associated with the ash or other distinctive elements of the fauna.

In all, thirty-five species, comprising more than half the total assemblage associated with the Pearlette volcanic ash, are widely distributed and are found both within glaciated territory and outside the glacial border. This wide distribution of many species, together with the fact that 20 per cent of the total number of species is restricted to the Pearlette faunal zone, gives the Pearlette assemblage distinctive qualities that serve to prove the essential contemporaneity of the sediments deposited in the zone. This confirms similar conclusions based on petrographic and stratigraphic data.

The distinctive features of the Pearlette faunal assemblage are summarized in figure 2.

### PALEOPHYSIGRAPHIC ASPECTS

The stratigraphic succession of stream-deposited sediments is of small value for regional correlation of Pleistocene units beyond the margins of the till sheets. The repeated episodes of continental glaciation gave rise to repeated successions of fluviatile sediments that resemble one another in many respects. In Nebraska these successions of sediments have been referred to as cycles (Schultz

and Stout, 1945, 1948; Condra, Reed, and Gordon, 1947; Lueninghoener, 1947) and treated as stratigraphic units, although each cycle contains named formations. In the lower Mississippi Valley, Fisk (1939, 1944) has described a comparable sequence of sediments, to which he applied formation names, one such formational unit being distinguished under each of the major Pleistocene terrace surfaces.

In the Great Plains major modifications of drainage during Pleistocene time seem to have prevented the development of orderly downward terrace steps such as occur in the lower Mississippi Valley. Thus, in this region, stratigraphic succession is not a usable tool for correlation over wide areas unless some special lithologic element, such as a volcanic ash deposit or a buried soil, is present in the cyclic sequence of sediments. Furthermore, paleophysiography is less usable in the Plains than in some other regions because locally younger sediments occur stratigraphically above older, whereas at other places within the Great Plains younger sediments occur on successively lower terraces, and at still other localities only part of the cyclic episodes are recorded. Paleophysiographic techniques are usable in local areas or in areas that possess a unified drainage history.

Perhaps the most important contribution paleophysiography can make toward the integration of Great Plains Pleistocene stratigraphy with the glacial section is the determination of the time relationship between fluvial and glacial cycles. Quite clearly, each cyclic episode of valley cut and fill is controlled directly by the advance and retreat of a continental ice sheet and by the associated climatic change, melt waters, fluctuations of sea-level, or isostatic depression and marginal superelevation of the earth's crust.

If, in the area peripheral to the ice, valley cutting was accomplished throughout interglacial times and filling throughout glacial times, the sediments serve as record only of the glacial episodes. If, on the other hand, valley cutting was accomplished during the glacial episodes, the sediments are a record of interglacial time. Still another possibility is that cutting of valleys took place during the advance and maximum stand of continental ice sheets (Ireland, 1946), alluviation of the valleys during the retreating phase of each glaciation, and that interglacial time was an interval of approximate equilibrium when soil making was the most important process. Several lines of evidence, including the nature of the fluvial sediments, relation of old soils to the several cycles of sedimentation, and physiographic history, suggest that the last of these three possibilities obtained in the region under investigation.

## STRATIGRAPHIC CONCLUSIONS

The establishment of the Pearlette volcanic ash zone as a datum of Yarmouthian age furnishes us with means for correlating the Great Plains Pleistocene section with the stratigraphic succession of the glaciated areas and for interrelation of events in the late Cenozoic history of the Plains with glacial chronology. Furthermore, it makes possible a major step toward unification of stratigraphic nomenclature for the cycle of sedimentation that contains the volcanic ash lentils. A diversity of stratigraphic names has been used for Pleistocene deposits in different parts of the Plains, and only in Nebraska has a terminology of state-wide scope been proposed. By utilizing terms adopted in Nebraska as member-names within the Sanborn formation (Frye and Fent, 1947), the nomenclature of the late Pleistocene deposits has been unified

throughout much of Kansas. It is here proposed to effect the same type of unification for the remaining post-Kansas till deposits of this region. Available data are not sufficient to allow similar treatment of the pre-Kansan Pleistocene deposits in the Great Plains region. Correlations and proposed nomenclatures are shown in figure 3.

### PROBLEMS OF THE BLANCAN

It is not the purpose of this paper to discuss the many problems that have arisen concerning placement of the Pliocene-Pleistocene boundary line. It is needful, however, to give attention to the cycle of fluvial sedimentation preceding the Kansan cycle in order to place deposition of the Pearlette zone in its proper perspective in Pleistocene history.

In the Nebraska area peripheral to the ice front, the cycle of coarse to fine sediments occurring stratigraphically below the "post-Kansas-till-sediments" has been classed as Holdrege and Fullerton formations (Lugn, 1935). These units have been correlated (Condra, Reed, and Gordon, 1947) with the Broadwater formation (Schultz and Stout, 1945), which occupies a high terrace position in western Nebraska and which has yielded an abundant fauna correlated with the Blancan (Wood, *et al.*, 1941) fauna of northwestern Texas. In Texas, where the type section of the Blanco and correlative deposits have been restudied recently (Evans and Meade, 1945; Meade, 1945), the deposits of Blancan age in Rita Blanca Canyon, west of Channing (loc. 46), occur below Pleistocene sediments that contain Pearlette ash and an associated snail fauna.

In central Kansas the pre-Kansan cycle of sediments occupies the bottom of a deep filled valley in Rice County where numerous test holes have revealed its presence below the strata of the Kansan cycle (including the Pearlette ash) and younger Pleistocene deposits (personal communication from O. S. Fent). This filled and abandoned valley was cut about 200 feet in bedrock after the deposition of the Pliocene Ogallala formation and before the deposition of the Blancan sediments (Frye and Fent, 1947, p. 46).

In southwestern Kansas the Rexroad formation occurs unconformably below the Meade formation, which contains the Pearlette ash and associated vertebrate and molluscan faunas. In this region, however, the relationships of the several units have been obscured partly by faulting and solution-subsidence. Furthermore, the inclusion (McLaughlin, 1946; Byrne and McLaughlin, 1948) in the Rexroad formation of beds that have yielded fossil vertebrates indicating an age older than type Rexroad fauna and the exclusion of other beds formerly mapped with Rexroad suggest that more than one pre-Kansan cycle has been included in the formation.

Data from these areas demonstrate the existence of a pre-Kansan cycle of alluvial sedimentation, associated with Nebraskan glaciation, that has been assigned to the Blancan provincial age (Wood, *et al.*, 1941; Elias, *et al.*, 1945) (although it may not represent all of Blancan time); this pre-Kansan body of sediments is separated from the Ogallala formation of Pliocene age by an important regional unconformity.

The Blancan faunas recently have been assigned by some workers (Schultz and Stout, 1941, 1945, 1948; McGrew, 1944; Evans and Meade, 1945) to early Pleistocene time and by others to late Pliocene time (Hibbard, 1937, 1941*a*, 1941*b;* G. G. Simpson, 1947; Frye and Hibbard, 1941*b*). Simpson (1947, p. 623) has correlated the Blancan with the Villafranchian and Astian of Europe, which are generally classed as Pliocene; accord-

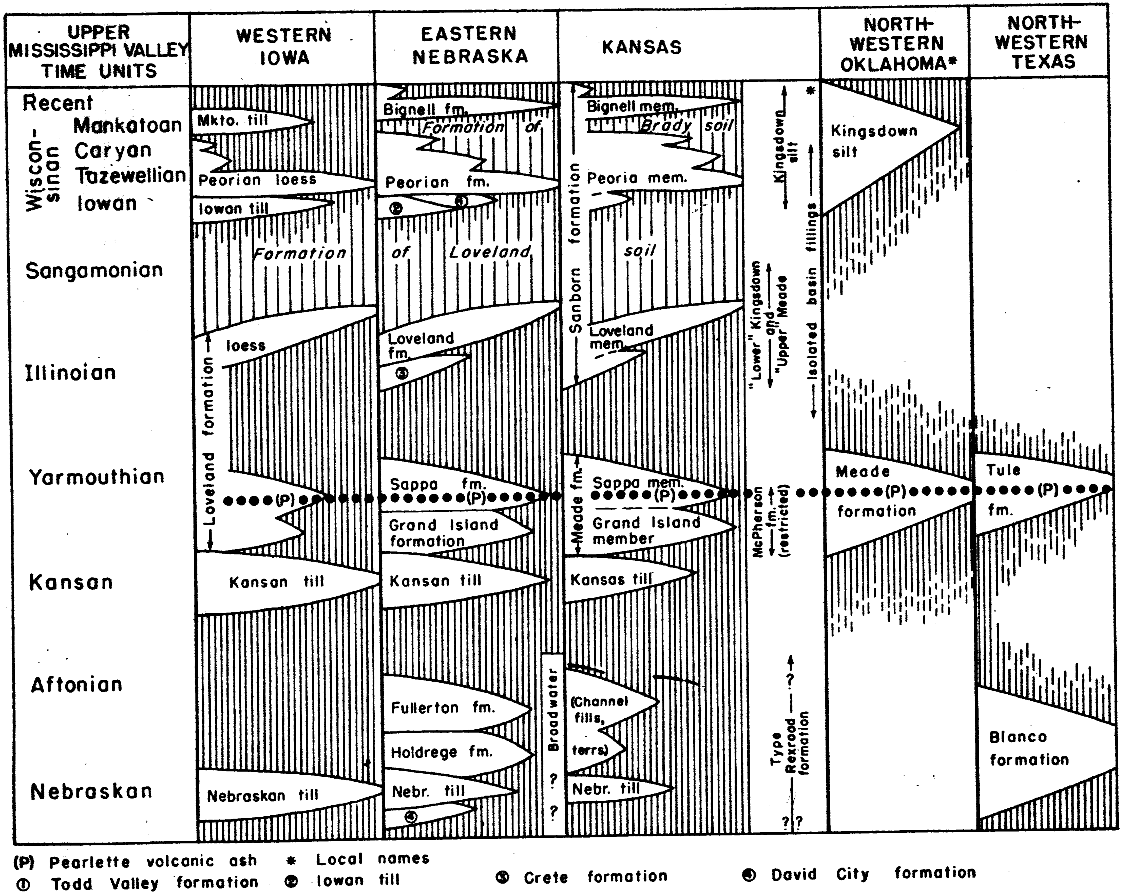



FIG. 3.—Chart showing classification and correlation of various stratigraphic units from the Missouri Valley area into the central Great Plains. The stratigraphic position of the Pearlette volcanic ash is shown by large dots.

ingly, he has interpreted the Blancan vertebrates as late Pliocene in age, but he did not correlate the Blancan deposits with the glacial sequence of north-central United States. Schultz and Stout (1948) suggest that the Blancan correlates with the Villafranchian of Europe and that the Plaisancian-Astian fauna may be represented in America by the fossil mammals of the Ash Hollow and Kimball (including Sidney) formations of Nebraska classification. The evidence presented here shows that the cycle of sedimentation represented by Holdrege-Fullerton (Broadwater) in Nebraska, the Blanco formation in Texas, the earliest channel fills of central Kansas, and the upper part, at least, of the Rexroad formation in southwestern Kansas are related to Nebraskan glaciation; and in terms of glacial chronology they should be placed in late Nebraskan and early Aftonian time. This correlation does not preclude their assignment to late Pliocene time if the beginning of Pleistocene time is interpreted to belong after the withdrawal of the first major ice sheet, following the view of some European workers (Gromov, 1945).

As the present study is concerned primarily with the Kansan cycle of sediments, which contains the Pearlette volcanic ash zone, no attempt is made here to modify the existing terminology of the pre-Kansan sediments. Figure 3 shows the names now in common use in several areas and their correlation with the generally accepted time scale developed in the upper Mississippi Valley.

### CLASSIFICATION IN KANSAS

It is here proposed to restrict the Meade formation (Frye and Hibbard, 1941*b*) to sediments at its type section, NW. ¼, Sec. 21, T. 33 S., R. 28 W., Meade County, Kansas, and equivalent deposits elsewhere, which signifies that this term should be used throughout Kansas to include the sediments of the Kansan cycle of fluvial sedimentation. Also, two members of the Meade formation are recognized, namely, the Grand Island member, below, and the Sappa member, above; the latter includes the Pearlette volcanic ash lentil. Frye and Fent (1947) have revised the classification of late Pleistocene deposits underlying extensive areas in the northern half and the south-central part of Kansas. The unification of terminology produced by this classification is indicated in the correlation chart, figure 3.

*Meade formation.*—The sediments of the Kansan cycle of alluviation constitute the best developed and most widely occurring Pleistocene stratigraphic unit, exclusive of the loesses, in the central Great Plains. Although the Blancan deposits have perhaps attracted more general notice among geologists because of their abundant fossil vertebrate fauna and controversial age, they occur only at scattered localities. The post-Meade cycles are less well developed and in some drainage basins hardly recognizable.

It was after the deposition of the Meade formation and correlative sediments that the most extensive modifications of the early Pleistocene drainage pattern occurred, and, as a result, these deposits are preserved in segments of former major valleys that no longer contain through-flowing streams. This early Pleistocene drainage, at least south of Nebraska, seemingly did not cross the Flint Hills area to join the Missouri-Mississippi system, as it now does, but flowed toward the south or southeast across Kansas and northern Oklahoma. Drainage from the western margin of the continental ice sheets and from the Rocky Mountain region converged west

of the Flint Hills area, and only after the retreat of the Kansan ice sheet was this Great Plains drainage diverted to the east. Notable among these abandoned valley segments are those of McPherson County and Lincoln and Ellsworth counties, Kansas. In McPherson County an abandoned valley (Lohman and Frye, 1940) extends from the present valley of the Smoky Hill River to the Arkansas River. The alluvial fill clearly displays sand and gravel (Grand Island member of the Meade formation) overlain conformably by stratified sand and silt which includes the Pearlette volcanic ash (Sappa member of the Meade formation), which is overlain in turn by Loveland silt of the Sanborn formation. The well-known fossil vertebrate fauna referred to as the "Equus beds" or McPherson fauna (Frye and Hibbard, 1941*a*) was collected from the Grand Island sand and gravel below the Pearlette volcanic ash. The abandoned-valley segment in Lincoln and Ellsworth counties, Kansas, has been called the Wilson Valley (Frye, Leonard, and Hibbard, 1943; Frye, 1945*b*) and extends across the present divide between the Saline Valley and the Smoky Hill Valley. Also, the succession of sediments in this area closely resembles that in the McPherson Valley of central Kansas; at Fullerton, Nebraska; near Little Sioux, Iowa; and at many other localities where this unit has been studied in exposures and in drill cuttings.

At the type locality of the Meade formation, sedimentation was in part controlled by solution-subsidence and related faulting, and the Sappa member, which contains the Pearlette volcanic ash of the type area (pl. 2, *A*), is unusually thick. It here contains beds of silty clay indicative of slack-water conditions. The Grand Island sand and gravel at the base of the type section nevertheless reflects a Rocky Mountain source and was probably deposited by a through-flowing stream trending toward the south-southeast.

At several localities in the High Plains, typically in Gove County, Kansas (loc. 20; pl. 1, *D*), and at several places in northwestern Texas (where this unit has been called the Tule formation) the Grand Island member consists largely or entirely of locally derived materials and presumably is a deposit of tributary streams that did not tap a glacial or mountain sedimentary source. This also is true at localities 25 and 26, within and on the east flank of the Flint Hills. Here this cycle is represented by deposits derived from the Permian bedrock of the region and contains the Pearlette volcanic ash in relatively coarse materials. The volcanic ash, which was not affected by drainage divides, and the snail fauna of the Sappa member, are nonetheless diagnostic.

The physiographic position of the Kansan cycle of sediments (Meade formation) varies in different areas owing to the differences in drainage history. In parts of eastern Nebraska (for example, near Fullerton) and western Iowa along the Missouri River bluffs, various Pleistocene sediments occur in stratigraphic sequence, the several units of cyclic sediments being set apart by unconformities. A similar stratigraphic succession is widely present in southwestern and parts of central Kansas, where the Pleistocene deposits fill deep valleys cut below the Ogallala surface or fill solution-subsidence or down-faulted areas.

Throughout much of the Kansas River Basin a physiographic sequence is distinguished, consisting of successively lower terraces, and this is comparable in some respects to the terrace succession of the lower Mississippi Valley described by Fisk (1939, 1944). The Meade formation

is particularly prominent along Smoky Hill River underlying the surface of a high terrace in Russell and Ellsworth counties, Kansas. In the panhandle region of Texas, deposits correlative with the Meade formation (Tule formation of Texas classification) occur stratigraphically above the Blanco formation or, where the Blanco is absent, above the Ogallala formation, and their upper surface is approximately accordant with the High Plains surface. In some places (loc. 48) the Tule formation is exposed at the crest of the High Plains escarpment, high above the lowland developed by the tributaries of the Red River system.

*Sanborn formation.*—In 1947 Frye and Fent reviewed usage of the Sanborn formation and expanded its geographic application to cover all of northern and central Kansas. It represents the time span of the Kingsdown silt of southwestern Kansas and correlative deposits of the panhandle regions of Oklahoma and Texas. The Sanborn formation includes beds classed by the Nebraska Geological Survey as Crete, Loveland, Todd Valley, Peorian, and Bignell formations and classed by the Iowa Geological Survey as Loveland loess and Peorian loess. In Kansas three members—Loveland silt, Peoria silt, and Bignell silt—which are separated by the Loveland and Brady (Schultz and Stout, 1948) soils, have been defined within the Sanborn formation. These members are stratigraphic equivalents to the comparably named units, which are classed as formations in Nebraska.

In Nebraska the deposits here included within the Sanborn formation are considered to represent several cycles of sedimentation that reflect the advance and retreat of Illinoian and the several Wisconsinan ice sheets (Schultz and Stout, 1945, 1948; Condra, Reed, and Gordon, 1947). Owing to the fact that these late Pleistocene ice invasions did not approach the central Great Plains region so closely as did the Kansan and Nebraskan, the cyclic repetition of valley cut and fill is much less well developed than is the case with the Nebraskan and Kansan cycles. Furthermore, the most widespread deposits assignable to the Sanborn are the loess sheets that mantle vast areas of the uplands and are, in many places, separated only by fossil soils. Therefore, the present classification that groups together the several late Pleistocene cycles into one formational unit seems most consistent with usable field practices in the central Great Plains.

Acknowledgments.—The manuscript for this paper was improved by the advice and criticisms of Raymond C. Moore, state geologist of Kansas. We express our thanks to the following persons who read and criticized the manuscript: H. G. Hershey, state geologist of Iowa, and A. C. Trowbridge of the University of Iowa; E. P. Rothrock, state geologist of South Dakota; G. E. Condra, state geologist, and E. C. Reed, associate state geologist of Nebraska; C. Bertrand Schultz, director, and T. M. Stout of the University of Nebraska State Museum; R. H. Dott, director, and S. L. Schoff of the Oklahoma Geological Survey; J. W. Stovall of the University of Oklahoma; J. T. Lonsdale, director, and G. L. Evans and G. E. Meade of the Texas Bureau of Economic Geology. Accompanying us on the May, 1947, field conference were, among those mentioned above, Trowbridge, Hershey, Stout, Schultz, and Reed; and, in addition, S. E. Harris, C. S. Gwynne, A. L. Lugn, W. D. Frankforter, and G. C. Lueninghoener. In August, 1947, Schultz, Stout, Lueninghoener, and Frankforter kindly guided us to the outcrops near Orleans, Nebraska. Richard F. Flint generously furnished us with data concerning the stratigraphy of the glacial deposits in the vicinity of locality 1 in southeastern South Dakota. During the early stages of field work on stratigraphy of the Pearlette volcanic ash in southwestern Kansas invaluable aid was given us by C. W. Hibbard, formerly curator of vertebrate paleontology at the University of Kansas Museum of Natural History.

## REFERENCES CITED


BYRNE, F. E., and MCLAUGHLIN, T. G. (1948) Geology and ground-water resources of Seward County, Kansas: Kansas Geol. Survey Bull. 69.

CONDRA, G. E.; REED, E. C.; and GORDON, E. D. (1947) Correlation of the Pleistocene deposits of Nebraska: Nebraska Geol. Survey Bull. 15, pp. 1–73.

CRAGIN, F. W. (1896) Preliminary notice of three late Neocene terranes: Colorado Coll. Studies, vol. 6, pp. 53–54.

ELIAS, M. K., *et al.* (1945) Blancan as a time term in the central Great Plains: Science, vol. 101, no. 2620, pp. 270–271.

EVANS, G. L., and MEADE, G. E. (1945) Quaternary of the Texas High Plains: Univ. Texas Pub. 4401, pp. 485–507.

FISK, H. N. (1939) Depositional terrace slopes in Louisiana: Jour. Geomorphology, vol. 2, pp. 181–200.

——— (1944) Geological investigation of the alluvial valley of the lower Mississippi River: War Dept., Corps of Engineers, Mississippi River Comm., Vicksburg, Mississippi.

FLINT, R. F. (1947) Glaciation of South Dakota (preliminary discussion) (abstr.): Geol. Soc. America Bull. 58, pp. 1179–1180.

FRYE, J. C. (1942) Geology and ground-water resources of Meade County, Kansas: Kansas Geol. Survey Bull. 45.

——— (1945*a*) Problems of Pleistocene stratigraphy in central and western Kansas: Jour. Geology, vol. 53, pp. 75–93.

——— (1945*b*) Valley erosion since Pliocene "algal limestone" deposition in central Kansas: Kansas Geol. Survey Bull. 60, pt. 3, pp. 85–100.

——— (1946) Review of studies of Pleistocene deposits in Kansas: Am. Jour. Sci., vol. 244, pp. 403–416.

——— and FENT, O. S. (1947) The late Pleistocene loesses of central Kansas: Kansas Geol. Survey Bull. 70, pt. 3, pp. 29–52.

——— and HIBBARD, C. W. (1941*a*) Stratigraphy and paleontology of a new Middle and Upper Pliocene formation of south-central Kansas: Jour. Geology, vol. 49, pp. 261–278.

——— — ——— (1941*b*) Pliocene and Pleistocene stratigraphy and paleontology of the Meade Basin, southwestern Kansas: Kansas Geol. Survey Bull. 38, pp. 389–424.

———; LEONARD, A. B.; and HIBBARD, C. W. (1943) Westward extension of the Kansas "Equus beds"; Jour. Geology, vol. 51, pp. 33–47.

GROMOV, V. (1945) Twenty-five years of study of the Quaternary in the U.S.S.R.: Am. Jour. Sci., vol. 243, pp. 492–516.

HIBBARD, C. W. (1937) An Upper Pliocene fauna from Meade County, Kansas: Kansas Acad. Sci. Trans., vol. 40, pp. 239–265.

——— (1941*a*) Paleoecology and correlation of the Rexroad fauna from the Upper Pliocene of southwestern Kansas, as indicated by the mammals: Kansas Univ. Sci. Bull. 27, pp. 79–104.

——— (1941*b*) Mammals of the Rexroad fauna from the Upper Pliocene of southwestern Kansas: Kansas Acad. Sci. Trans., vol. 44, pp. 265–313.

——— (1944) Stratigraphy and vertebrate paleontology of Pleistocene deposits of southwestern Kansas: Geol. Soc. America Bull. 55, pp. 707–754.

———; FRYE, J. C.; and LEONARD, A. B. (1944) Reconnaissance of Pleistocene deposits in north-central Kansas: Kansas Geol. Survey Bull. 52, pp. 1–28.

IRELAND, H. A. (1946) Cause and age of Pleistocene entrenchment in the Ohio and Lake Erie basins (abstr.): Geol. Soc. America Bull. 57, pp. 1206–1207.

KAY, G. F. (1924) Recent studies of the Pleistocene in western Iowa (abstr.): Geol. Soc. America Bull. 35, pp. 71–73.

——— and APFEL, E. T. (1928) The pre-Illinoian Pleistocene geology of Iowa: Iowa Geol. Survey, vol. 34, pp. 1–304.

——— and GRAHAM, J. B. (1943) The Illinoian and post-Illinoian Pleistocene geology of Iowa: Iowa Geol. Survey, vol. 38, pp. 11–262.

LANDES, K. K. (1928) Volcanic ash resources of Kansas: Kansas Geol. Survey Bull. 14.

LEONARD, A. B., and FRANZEN, D. S. (1944) Mollusca of the Laverne formation (Lower Pliocene) of Beaver County, Oklahoma: Kansas Univ. Sci. Bull. 30, pp. 15–39.

——— and FRYE, J. C. (1943) Additional studies of the Sanborn formation, Pleistocene, in northwestern Kansas: Am. Jour. Sci., vol. 241, pp. 453–462.

LOHMAN, S. W., and FRYE, J. C. (1940) Geology and ground-water resources of the "Equus beds" area in south-central Kansas: Econ. Geology, vol. 35, pp. 839–866.

LUENINGHOENER, G. C. (1947) The post-Kansan geologic history of the Lower Platte Valley area: Univ. Nebraska Studies, new ser., no. 2.

LUGN, A. L. (1935) The Pleistocene geology of Nebraska: Nebraska Geol. Survey Bull. 10, pp. 1–223.

MCGREW, P. O. (1944) An early Pleistocene (Blancan) fauna from Nebraska: Field Mus. Nat. History Pub., Geol. ser., vol. 9, no. 2, pp. 33–66.

MCLAUGHLIN, T. G. (1946) Geology and ground-water resources of Grant, Haskell, and Stevens counties, Kansas: Kansas Geol. Survey Bull. 61.

MEADE, G. E. (1945) The Blanco fauna: Univ. Texas Pub. 4401, pp. 509–556.

RIDGWAY, ROBERT (1912) Color standards and color nomenclature, Washington, D.C.

SCHULTZ, C. B., and STOUT, T. M. (1941) Guide for a field conference on the Tertiary and Pleistocene of Nebraska (in collaboration with A. L. Lugn, M. K. Elias, F. W. Johnson, and

M. F. Skinner), first field conference of the Society of Vertebrate Paleontology, 1941; Univ. Nebraska State Mus., Special Pub., pp. 1–15.

——— — ——— (1945) Pleistocene loess deposits of Nebraska: Am. Jour. Sci., vol. 243, pp. 231–244.

——— — ——— (1948) Pleistocene mammals and terraces in the Great Plains: Geol. Soc. America Bull. (in press).

SIDWELL, RAYMOND, and BRONAUGH, R. L. (1946) Volcanic sediments in north Texas: Jour. Sedimentary Petrology, vol. 16, pp. 15–18.

SIMPSON, G. G. (1947) Holarctic mammalian faunas and continental relationships during the Cenozoic: Geol. Soc. America Bull. 58, pp. 613–688.

SIMPSON, H. E. (1947) Pleistocene geology of the Yankton area, South Dakota and Nebraska (abstr.): Geol. Soc. America Bull. 58, pp. 1228–1229.

SMITH, G. D., and RIECKEN, F. F. (1947) The Iowan drift border of northwestern Iowa: Am. Jour. Sci., vol. 245, pp. 706–713.

SWINEFORD, ADA, and FRYE, J. C. (1946) Petrographic comparison of Pliocene and Pleistocene volcanic ash from western Kansas: Kansas Geol. Survey Bull. 64, pt. 1, pp. 1–32.

TODD, J. E. (1892) Volcanic dust from Omaha, Nebraska: Am. Geologist, vol. 10, pp. 295–296.

——— (1895) Volcanic ash bed near Omaha: *ibid.*, vol. 15, p. 130.

——— (1899) The moraines of southeastern South Dakota and their attendant deposits: U.S. Geol. Survey Bull. 158.

WARREN, C. R. (1947) Iowan till at Chamberlain, South Dakota (abstr.): Geol. Soc. America Bull. 58, p. 1237.

WOOD, H. E., *et al.* (1941) Nomenclature and correlation of the North American continental Tertiary: Geol. Soc. America Bull. 52, pp. 1–48.

# BLACK HILLS TERRACE GRAVELS: A STUDY IN SEDIMENT TRANSPORT[1]

WILLIAM J. PLUMLEY
University of Chicago

ABSTRACT

Terraces were mapped along three streams which flow eastward from the Black Hills of South Dakota. Five erosional surfaces were found which represent major pauses in the downcutting activity of the regional drainage. The presence of knickpoints in the present-day stream profiles offers proof that the more recent terraces of this region are the result of factors controlling downcutting downstream in the Cheyenne and Missouri rivers. These factors are best explained on the basis of Pleistocene glaciation in eastern South Dakota.

Twenty-three channel samples were taken from the terraces of Bear Butte, Rapid, and Battle creeks. In addition, sand samples 1–2 mm. in size were collected from the headwaters of Battle Creek near Harney Peak to the lower end of the Cheyenne River where it flows into the Missouri River, a distance of about 200 miles. The samples were analyzed for size, shape, roundness, and lithology. Systematic changes as a function of distance were observed for mean size, skewness, lithology, shape, and roundness. The rate of rounding was found to be directly proportional not only to the difference between the roundness at some point and a limiting roundness but also to some power of the distance transported.

An attempt is made to express quantitatively the effects of stream transport on roundness and lithology in respect to "indices of maturity."

## Introduction

The field of sedimentology has rapidly advanced with the advent of modern methods of sediment analysis. Emphasis has shifted from the purely descriptive aspects of the science to an analytic approach. With this change in viewpoint the fundamental attributes of sedimentary particles have received much attention. These attributes, such as size, shape, and roundness, to name but a few, have been extensively studied by mathematical and statistical methods. The result of such methods has been to furnish a large amount of data which again tend to be merely descriptive. Far more important than the fact that a gravel deposit has a specific type of size-frequency distribution is the statement of what genetic factors were involved in producing that distribution.

The statement of these genetic factors involves a study of environments of deposition and, in addition, knowledge of the previous history of the material being deposited. This knowledge has too often been derived from theoretical reasoning rather than from experimental fact. The present research has been conducted with the purpose of adding to this knowledge by means of field observations.

The fundamental attributes of sediments are affected to a greater or less degree by transportation in streams. The present work is concerned with the effects of transportation on the attributes of size, shape, roundness, and lithology.

The work is presented in two parts. The first considers the physiographic expression and development of certain stream terraces within and east of the Black Hills of South Dakota. The second part is concerned with the sampling and analysis of material from these terraces.

## Physiography

### GEOLOGIC SETTING

The area of study includes the eastern border of the Black Hills of South Dakota and adjacent portions of the Missouri Plateau section of the Great Plains (fig. 1).

[1] Manuscript received July 30, 1948.

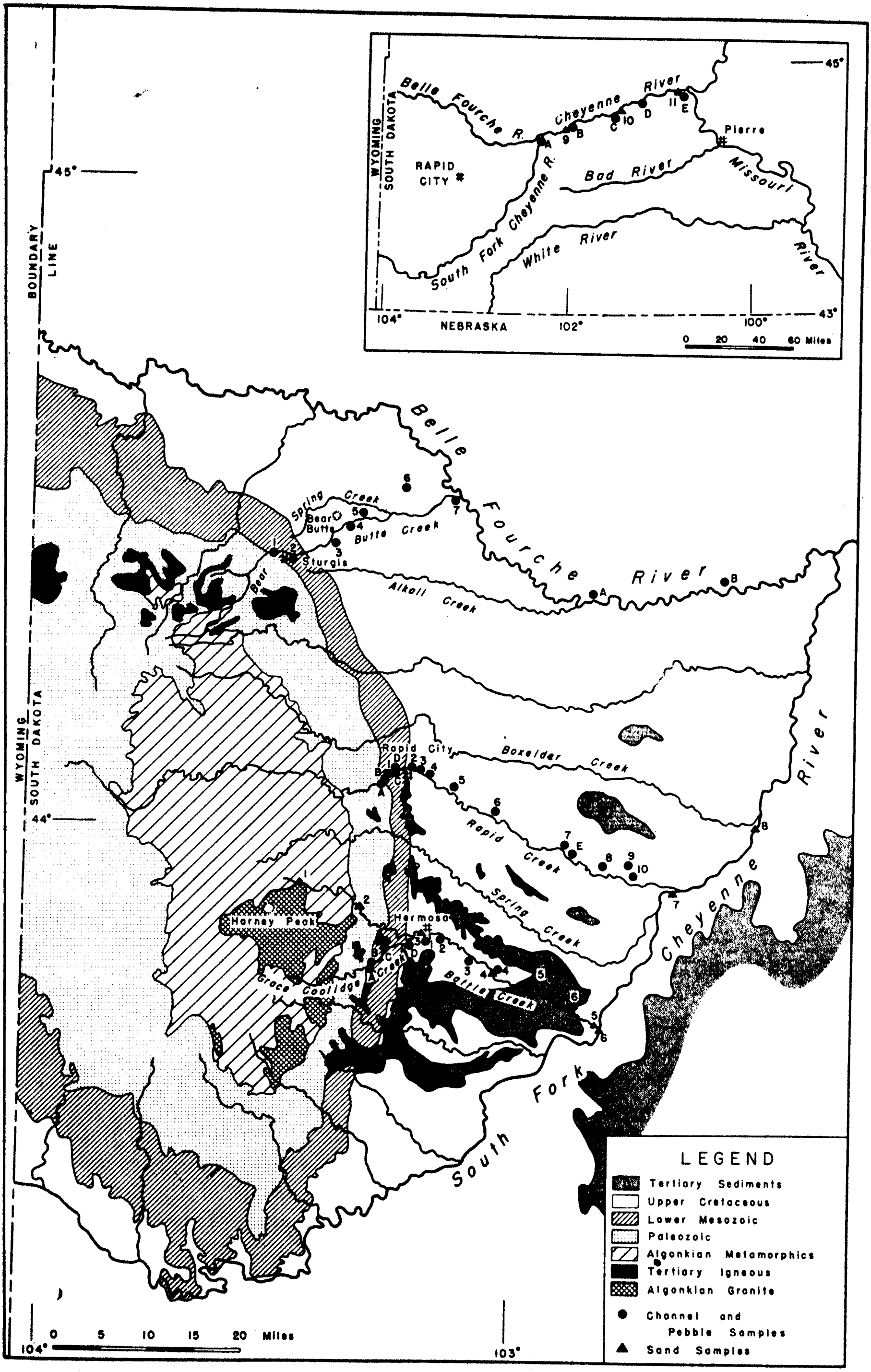



FIG. 1.—Index map of area studied

The Black Hills are an ovate mountainous area of moderate relief underlain by pre-Cambrian crystalline rocks and flanked by tilted sedimentary formations ranging in age from Cambrian to Cretaceous. Tertiary intrusives are found in the northern Hills, and flat-lying Tertiary sediments overlap Mesozoic formations on the plains to the east.

The Black Hills lie entirely within the drainage basin of the Cheyenne River and are encircled by two major tributaries, the Belle Fourche River on the north and the Cheyenne South Fork on the south. From the junction of these tributaries east of the Hills, the Cheyenne River flows eastward to enter the Missouri River north of Pierre, South Dakota.

Numerous secondary tributaries of the Belle Fourche and the Cheyenne South Fork flow eastward from the Hills. Detailed studies were made of deposits of three of these tributaries: (1) Bear Butte Creek, which drains the Tertiary intrusive area of the northern Hills, (2) Rapid Creek, which drains the metamorphic area in the central Hills, (3) Battle Creek, which drains the southern granitic area.

## PREVIOUS STUDIES

Prior to 1929 few attempts had been made to unravel the Cenozoic history of the area. The occurrence of high-level Tertiary gravels within the Hills and on the plains to the east was noted by Newton and Jenney (1880, p. 44), Crosby (1882, pp. 516–517), Todd (1900, pp. 120–121), and Darton (1909, pp. 58–59). Newton, Todd, and Darton also mention the presence of Quaternary terraces within the eastern foothill belt and along the tributaries of the Cheyenne River. These accounts are limited to short descriptions of the various terraces, with no attempt at regional interpretation.

Fillman (1929, pp. 1–48) attempted to integrate these terraces into a logical history of development. Three gravel-covered terraces were recognized and named: Mountain Meadow (Tertiary), Rapid (Quaternary), and Sturgis (Quaternary).

The Mountain Meadow surface is described by Fillman as having a late mature topography with local relief 1,000–1,500 feet within the Hills. The type locality is the Mountain Meadow upland, 1½ miles east of Deadwood. The Mountain Meadow surface is widely distributed in the interior basin and occurs as flat-topped divides on the interstream areas of the adjacent plains east of the Hills. Mountain Meadow time was believed to be brought to a close in mid-Oligocene by differential uplift, which amounted to 2,000–3,000 feet in the central Hills, 250 feet in the foothills, and 20–30 feet in the Big Badlands to the east. This uplift resulted in dissection of the Mountain Meadow surface to form the present valleys within the Hills. At the same time a peneplain was believed to have developed on the plains to the east.

The name "Rapid" was given to the highest gravel-capped terraces along the valleys within the foothills. The type locality is Rapid Creek Valley. Fillman (1929, p. 37) describes the relation between the Mountain Meadow surface and the Rapid terrace as follows: "When the so-called Upper Terrace (Rapid) is traced from the Hills proper it is found to cut across the Mountain Meadow surface on the Plains." As measured by Fillman, the height of the Rapid terrace above drainage on the border of the Hills was found to be 100 ± 5 feet. At the end of Rapid cycle time, the streams began to degrade their valleys once more, leaving the Rapid deposits exposed as gravel-capped terraces. This resumption of

stream downcutting is thought by Fillman to be related to the glaciation of eastern South Dakota or to other climatic changes.

The name "Sturgis" was given to those terraces which are found below the Rapid terrace in all the streams draining the Hills. The town of Sturgis is the type locality. The Sturgis terrace, as measured by Fillman, was found to occur about 50 feet above the present flood plains and consequently about 50 feet below the Rapid terrace level.

### FIELD WORK

The field work was greatly facilitated by use of aerial photographs. The photos were examined stereoscopically, and all visible terrace and flat high-level surfaces were outlined. The terrace outlines were verified by reconnaissance. Profiles of Bear Butte, Rapid, and Battle creeks and their terraces were made by means of a Wallace and Tiernen Sensitive Altimeter. The profile of each terrace was constructed with the stream profile as reference level. The final terrace maps were prepared from the aerial photographs by means of the slotted-template method (Kelsh, 1940, pp. 1–49) and vertical Sketchmaster.

Analysis of the field data reveals certain fundamental differences between the physiographic interpretation presented by Fillman and that of the present work. These differences are described in detail in the following section.

### DESCRIPTION OF TERRACE SYSTEMS

#### GENERAL STATEMENT

The most striking feature of the streams flowing eastward out of the Black Hills is the asymmetrical form of their north-south profiles. This feature is readily seen on the topographic maps of the Rapid and Hermosa quadrangles. The southern slopes of the majority of the creek valleys are very steep compared to the gentle northern slopes. Bear Butte and Rapid creeks possess this attribute to a high degree, and, as a result, terraces are commonly found on the northern sides of their valleys.

To avoid confusion in nomenclature, the names given to various surfaces by Fillman have been kept, even though considerable revision of heights and locations of these surfaces is necessary. In addition to the Mountain Meadow, Rapid, and Sturgis surfaces, two other terraces were found.

#### RAPID CREEK

In figure 2 four erosional surfaces are mapped. The outlines of these surfaces represent the relatively flat areas only. The highest surface is the Mountain Meadow, which is found topping the divide between Rapid Creek and Boxelder Creek to the north. At the first two elevation points on the Rapid Creek profile (fig. 3) this high surface occurs as a flat area whose outlines can be mapped. The two elevation points farther to the east represent the highest points on the divide between Rapid and Boxelder creeks. These latter points are considered to be below the Mountain Meadow surface, because joining them with the first two elevation points forms a convex profile, which means that subsequent erosion has removed the Mountain Meadow surface from the drainage divide a few miles east of the Hills.

The next lower surface is the Rapid terrace, which in the vicinity of Rapid City is at an elevation of 180 feet above Rapid Creek. This elevation is 80 feet higher than the one given by Fillman. One remnant of this terrace is found in

the Red Valley west of Rapid City. The next remnant to the east forms an elevated and steep-sided flat within Rapid City. As it is traced eastward, the Rapid terrace becomes wider, and its boundaries away from the stream are less distinct. East of Rapid City it is found exclusively on the north side of Rapid Creek. Nevertheless, Fillman mapped the Rapid terrace on both sides of These terraces are located between Dry Creek and Rapid Creek and between Antelope Creek and Rapid Creek. Reference to the profile (fig. 3) reveals that the gradients of these terraces are related neither to the gradient of Rapid Creek nor to those of its terrace systems. On the contrary, it is probable that these terraces represent former stream flats of Dry and Antelope creeks. The lower

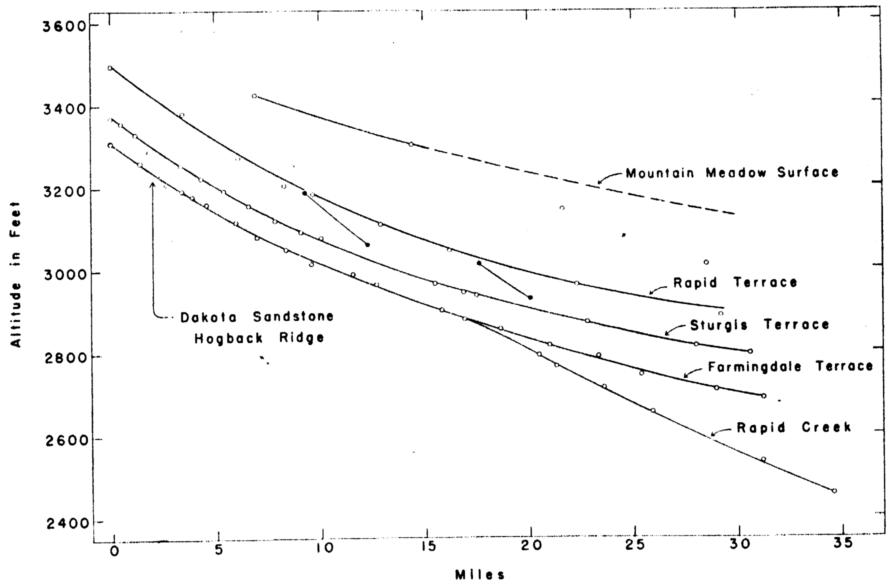



FIG. 3.—Rapid Creek profile

Rapid Creek. The terrace remnants south of Rapid Creek, however, are related to creeks tributary to Rapid Creek and not to Rapid Creek itself.

The Sturgis terrace is present throughout Rapid Creek Valley. Near Rapid City it is about 60 feet above drainage. Like the Rapid terrace, it is found predominantly on the north side of the valley. Two conspicuous terrace systems are found on the south side of the valley, which are related to the Sturgis cycle but are not at the same elevations.

end of each of these terraces appears to merge with the Sturgis terrace of Rapid Creek. Therefore, they probably represent the levels of Dry and Antelope creeks during Sturgis cycle time. These terraces which border Rapid Creek on the south and yet are not directly related to it are evidence that at the time of their formation Rapid Creek must have been located farther north than it now is. At that time a divide must have existed between Rapid Creek and the two tributaries. Later southward migration of

Rapid Creek removed all trace of that divide, with the exception of one remnant of higher ground between Rapid Creek and the terrace of Dry Creek. In figure 4 these relations are shown diagrammatically. The block diagram pictures the condition existing during the Sturgis cycle. The dashed lines represent the present conditions resulting from southward migration of Rapid Creek in post-Sturgis time.

A fourth, hitherto unknown, terrace system was found in Rapid Creek Valley. It is named the "Farmingdale terrace" because of its inception near the Farmingdale railroad stop. It appears in Rapid Creek Valley only between Farmingdale and the Cheyenne River. West of Farmingdale it coincides with the present-day flood plain of Rapid Creek. Reference to the profile (fig. 3) reveals that this terrace owes its presence to a marked increase in the gradient of Rapid Creek east of Farmingdale. The Farmingdale terrace is conspicuous along the Cheyenne River.

One other terrace should be mentioned to complete the record. It lies about 2 miles southeast of Rapid City and trends north-south at an elevation intermediate between the Rapid and the Sturgis terraces. It was not possible on the available field evidence to correlate this terrace with those of the other creeks studied.

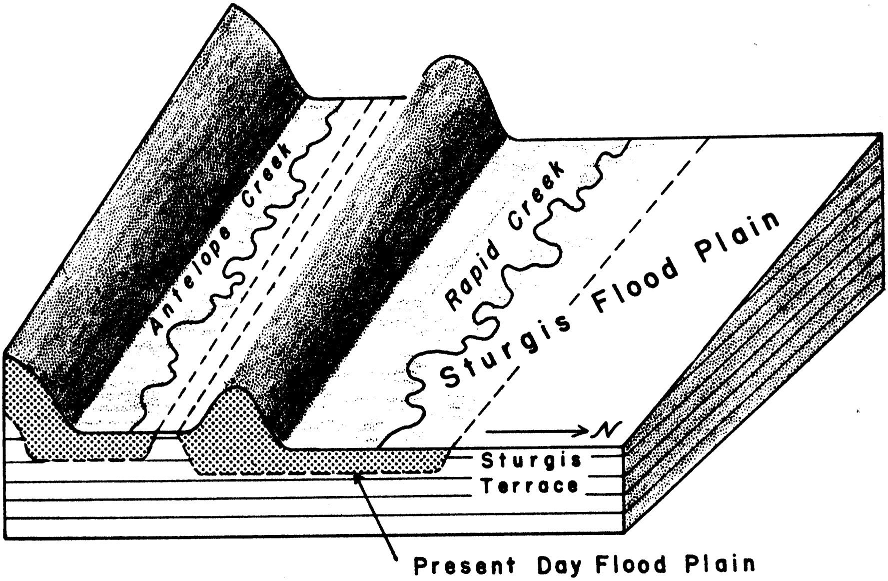



FIG. 4.—Diagrammatic sketch of probable land surface in Rapid Creek Valley during the Sturgis cycle

BEAR BUTTE CREEK

The terraces of Bear Butte Creek, with a few important exceptions, are similar to those of Rapid Creek. The Rapid and Sturgis terraces are present, but the Mountain Meadow surface is absent. In addition, there are two terraces below the Sturgis level (figs. 5 and 6).

The highest level is the Rapid terrace, which in the vicinity of the town of

Sturgis is 170 feet above drainage (according to Fillman, 100 feet). Unlike the Rapid terrace along Rapid Creek, which continues as a distinct terrace many miles east of the Hills, the Rapid terrace along Bear Butte Creek occupies an interstream divide a few miles east of Sturgis. As may be seen in figure 5, the Rapid surface occupies the highest land cause of the asymmetry of the valley, however, this plains surface is considerably below the Rapid surface south of the creek. Fillman mapped this area north of Bear Butte Creek as the Mountain Meadow surface. However, the Sturgis terrace profile shows this plains area to be merely a continuation of the Sturgis terrace.

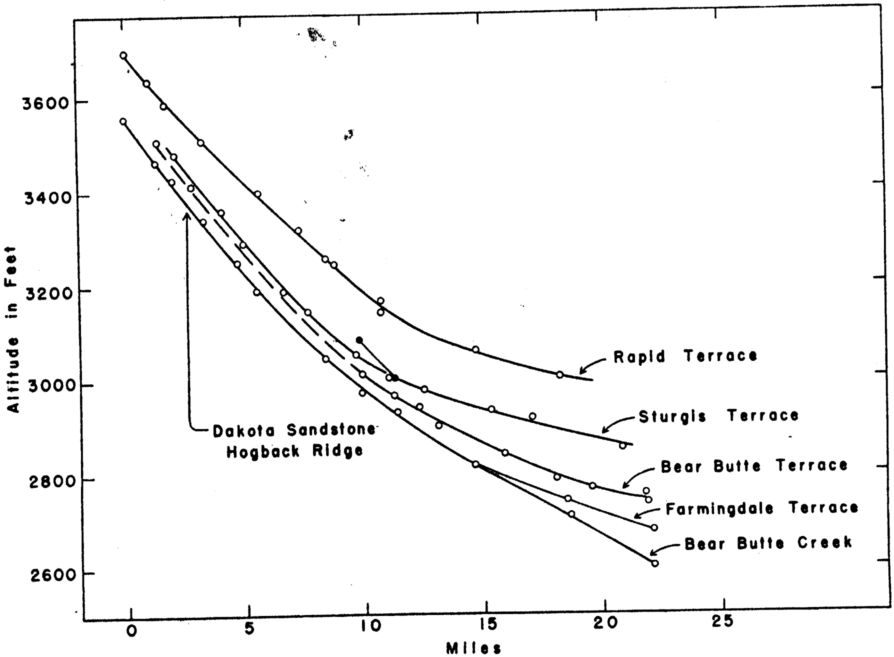



FIG. 6.—Bear Butte Creek profile

between Bear Butte Creek and Alkali Creek to the south. Fillman mapped this surface as Mountain Meadow. The profile (fig. 6) shows this to be incorrect as this stream-divide area is a direct continuation of the Rapid terrace closer to the Hills.

Within the foothill belt near the town of Sturgis, the Sturgis terrace occurs 70 feet above drainage. East of Bear Butte the Sturgis terrace merges with the plains surface north of Bear Butte Creek. Because

Two terraces occur below the Sturgis. The highest of these occurs as a major terrace along the Belle Fourche River. It is prominent along Bear Butte Creek as far west as Bear Butte. A few possible remnants are detectable near Sturgis. This terrace is here named the "Bear Butte terrace."

Below the Bear Butte terrace lies a fourth terrace, which is confined to the lower portion of the valley and cannot be traced west of a knickpoint in the stream,

rraces

at which place it appears to merge with the present-day flood plain. It is tentatively correlated with the Farmingdale terrace of Rapid Creek Valley.

An anomalous terrace occurs between Bear Butte and Bear Butte Creek. Its eastern extremity merges with the Sturgis terrace between Bear Butte and Spring creeks. Its western end merges with the talus slopes of Bear Butte. This surface is apparently a remnant of a previously more extensive alluvial fan of Bear Butte.

### BATTLE CREEK

The Rapid cycle is represented in Battle Creek Valley by both terraces and interstream divides. Remnants of the Rapid terrace can be traced several miles west of Hermosa up the valley of Grace Coolidge Creek (figs. 7 and 8). Near the town of Hermosa the Rapid terrace occurs 230 feet above drainage. Little correlation was noted between its location as reported by Fillman and as found by the writer. South of Battle Creek the Rapid terrace appears as a bench extending about 1½ miles east of Hermosa. East of this point it gives way to higher ground, capped by resistant Tertiary conglomerate. About 10 miles southeast of Hermosa the Rapid terrace is found on the north side of the valley. In this area the Rapid surface forms the divide between Battle Creek and Spring Creek to the north.

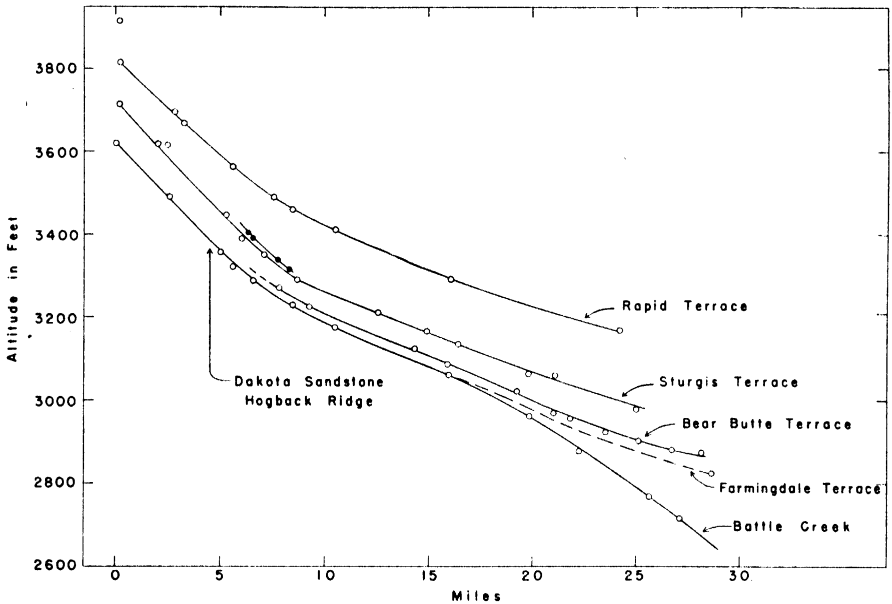



FIG. 8.—Battle Creek and Grace Coolidge Creek profile

Remnants of the Sturgis terrace extend several miles up Grace Coolidge Creek. It can be traced eastward from Hermosa to the Cheyenne River.

Thirteen miles southeast of Hermosa the gradient of Battle Creek suddenly steepens (fig. 8). Upstream from this knickpoint a low terrace is traceable westward almost to Hermosa. In places

this terrace almost fills the valley. The terrace extends downstream from the knickpoint and is widespread in the Cheyenne River Valley. It is tentatively correlated with the Bear Butte terrace of Bear Butte Creek Valley.

The Farmingdale terrace is absent along Battle Creek. It is present, however, along the Cheyenne River near the mouth of Battle Creek.

Two anomalous surfaces appear along Battle Creek Valley. In T. 3 S., R. 9 E., S. 28 a gravel-capped butte was found at an elevation intermediate between the Rapid and the Sturgis terraces. It is capped by about 6 feet of gravel lying upon a layer of Tertiary conglomerate. Perhaps during the Sturgis cycle, downcutting was retarded when the level of this resistant conglomerate was reached. Battle Creek later cut down through this hard layer and has since removed all trace of it, with the exception of one gravel-capped remnant.

On the north side of Battle Creek, east of Hermosa, is a large, relatively flat area intermediate in height between the Rapid and the Sturgis terraces. Gravel was observed on this surface near elevation point 3,405. The writer is unable to state without further evidence whether this surface has regional significance or is peculiar to the Battle Creek area only.

At the head of Grace Coolidge Creek an elongated gravel-capped flat was found about 100 feet above the Rapid terrace. This surface may possibly be the Mountain Meadow.

## HISTORY OF THE TERRACE CYCLES

During and following the uplift of the Black Hills in the Laramide revolution the area was subjected to erosion. No Eocene deposits are present in the area; either they were never deposited or they were removed prior to Oligocene deposition. The Oligocene White River group (Chadron and Brule formations) is a fluviatile and lacustrine deposit. The presence of basal conglomerates in the Chadron formation (Lower White River) indicates moderate relief in the Hills in early Oligocene time. The generally fine texture of the overlying White River deposits indicates deposition by streams of gentle gradient and in local lakes or bayous.

According to Darton (1899–1900, p. 561), the Hills were uplifted several hundred feet subsequent to White River deposition. Material from the Hills contributed to the coarser fluviatile deposits of Miocene age which cap the White River deposits south of the White River. Wanless (1923, pp. 190–269) considers Miocene beds capping White River beds on isolated buttes northeast of the Hills evidence that Miocene sediments extended east and northeast of the Black Hills as well as southeast. He concludes that a period of peneplanation followed Miocene or Pliocene deposition, in which beds of the same age were eroded more in South Dakota than in Nebraska. Remnants of this surface (the High Plains) exist as table tops and cut across Pierre shale (Cretaceous) on the north and Miocene sediments on the south. Fenneman (1931, pp. 61–79) states that Pliocene sediments (Ogallala formation) were being deposited in Nebraska at the same time that erosion was proceeding to the north in South Dakota. He therefore dates the High Plains surface as Pliocene. Degradation continued subsequent to the formation of this erosion surface in South Dakota, so that today it exists only as isolated flat-topped buttes north of the Pine Ridge escarpment.

These regional features must be considered in discussing the age of the Mountain Meadow surface. Fillman

(1929, pp. 1–48) contends that the Mountain Meadow surface was produced in mid-Oligocene time, following deposition of the Chadron formation. She states that, before deposition of the Brule formation (Upper Oligocene), the Mountain Meadow surface was uplifted 2,000–3,000 feet in the Hills and 20–30 feet in the Big Badlands. Her evidence for the age of this surface is (1) the mid-Oligocene age of the fossils contained in the gravels which lie upon it and (2) the fact that the Mountain Meadow surface plunges beneath younger materials (Rapid terrace gravels—Pleistocene) a few miles east of the Hills. It is thought to be correlative with the Chadron-Brule unconformity in the Big Badlands.

The writer's field observations are at variance with Fillman's hypothesis for the following reasons: (1) The Mountain Meadow surface as mapped by Fillman represents several cycles of erosion. For example, the stream-divide surface between Bear Butte and Alkali creeks is correlative with the Rapid terrace to the west. The plains surface between Bear Butte Creek and the Belle Fourche River in T. 7 N. and R. 7 E. is correlative with the Sturgis terrace to the west. East of Hermosa large areas of the upland between Battle and Spring creeks have been reduced to lower elevations since the formation of the Rapid surface in the vicinity. Each of these surfaces has been mapped as Mountain Meadow by Fillman. Thus the term "Mountain Meadow surface" loses its genetic significance when applied to the plains area east of the Hills unless its limits can be more rigorously outlined. (2) East of Rapid City the divide between Rapid and Boxelder creeks is flat-topped and gravel-covered. The writer believes that this high surface is correctly identified as the Mountain Meadow surface and is correlative with the subsummit surface of that name in the Hills. The west-east profile of this surface does not intersect the Rapid surface to the east, as contended by Fillman. Therefore, any correlation of the Mountain Meadow surface east of Rapid City with the mid-Oligocene unconformity in the Big Badlands is extremely doubtful.

Darton's report (1899–1900, p. 543) reveals that mid-Oligocene fossils were found in White River deposits high in the Hills southwest of Hermosa. However, this proves nothing about Mountain Meadow age unless these beds can be established as equivalent to gravels capping the Mountain Meadow surface. On the basis of the discrepancies indicated in 1 above, this correlation must be suspect.

The above analysis indicates that neither fossil nor physiographic evidence dates the Mountain Meadow surface as mid-Oligocene. Meyerhoff and Olmstead (1937, p. 306) report that the subsummit erosional surface in the Black Hills was formed contemporaneously with the Miocene depositional surface of the High Plains in southeastern Wyoming and Nebraska (as noted previously, Fenneman assigns a Pliocene age to the High Plains). They point out that earlier studies have argued for an Eocene age because White River sediments are found in valleys at elevations distinctly lower than the subsummit level. They contend, however, that the distribution and variable elevations of the White River deposits suggest valley-filling in a more rugged and therefore more ancient country than that represented in the subdued topography of the Black Hills. They conclude, therefore, that the valleys were filled and the interstream divides peneplained simultaneously and that equilibrium between the complementary

processes of degradation and aggradation was achieved in Miocene time.

From the foregoing considerations it is possible to conclude only that the true Mountain Meadow surface, as exemplified by the subsummit surface within the Hills and the Rapid–Boxelder Creek divide east of the Hills, was produced by degradation during Miocene or Pliocene time.

Subsequent to the formation of the Mountain Meadow surface a new cycle of erosion was begun. Deep valleys were excavated within the Hills, and a peneplain surface was developed in the area to the east (Rapid cycle). Close to the Hills the planation was least effective, as shown by the Mountain Meadow surface still preserved on the stream divides.

TABLE 1

CORRELATION OF STREAM TERRACES IN THE BLACK HILLS REGION

| Battle Creek | Rapid Creek | Bear Butte Creek |
|---|---|---|
| Rapid terrace | Rapid terrace | Rapid terrace |
| Sturgis terrace | Sturgis terrace | Sturgis terrace |
| Bear Butte terrace | | Bear Butte terrace |
| | Farmingdale terrace | Farmingdale terrace |

During Rapid cycle time the streams flowing east from the Hills did not enter the Cheyenne South Fork as they do today. Todd (1902, pp. 27–40) noted that numerous Black Hills erratics capped extensive areas in the valleys of the White and Bad rivers (inset map, fig. 1) and that shallow, gravel-filled channels crossed the Cheyenne–White River divide, which is at an elevation comparable to that of the Rapid surface to the west. Therefore, Todd concluded that in Pliocene time drainage was eastward from the Black Hills across the present South Fork of the Cheyenne River and that also in later Pliocene time a southward-working tributary of the Cheyenne River captured the headwaters of the White and Bad rivers. Perisho and Visher (1912, pp. 57–58) and Wanless (1923, pp. 190–269) concurred with Todd's findings but placed this series of events in the Pleistocene. This latter conclusion is supported by Fillman's report on the discovery of a Pleistocene horse's tooth in the Rapid terrace gravels.

Following the widespread planation of the Rapid cycle, the streams again resumed downcutting in the Sturgis cycle of erosion. Before the graded condition was reached in this cycle, streams such as Rapid and Battle creeks were integrated into one system by the pirating action of the Cheyenne South Fork. The Sturgis terrace levels are well below the Cheyenne–White River divide, thus placing the time of piracy after Rapid cycle time and before completion of the Sturgis cycle.

Analysis of the physiographic history from the close of the Sturgis cycle to the present is complicated by two other terrace systems which reflect pauses in the reduction of the land. Although two distinct terrace levels below the Sturgis level are present in the valleys of the Belle Fourche and Cheyenne South Fork, both levels are not represented in Rapid, Bear Butte, and Battle Creek valleys. Table 1 lists the terraces present in each valley and their possible correlation.

This correlation is based primarily on the number and the topographic expression of terraces within the foothill belt. Two terraces are present within this belt in all the creeks studied. The lower terrace (Sturgis) rises abruptly from the present-day flood plain, and the upper terrace (Rapid) rises abruptly from the lower terrace. Thus a correlation of these two sets of terraces from one valley to

another is believed to be valid. The fact that these two sets of terraces are not at the same elevation in each valley does not invalidate the correlation. Because the creeks under consideration at one time flowed eastward, independently, across western South Dakota (prior to capture by the Cheyenne South Fork), the factors influencing rate of downcutting may well have varied in the different creeks.

The correlation of terraces below the Sturgis terrace is based on the assumption that the Bear Butte terrace was once present in Rapid Creek Valley but has since been completely removed in the planation accompanying the Farmingdale cycle. This event is quite possible, since the Bear Butte terrace is almost completely missing in the upper half of the smaller Bear Butte Creek Valley. It follows logically that the larger stream of Rapid Creek would be capable of completely removing the Bear Butte terrace.

The Farmingdale cycle of erosion is represented in Battle Creek by the present-day flood plain upstream from the Battle Creek knickpoint. Downstream from this point the Farmingdale terrace has been completely removed by stream side-cutting of the present cycle.

Knickpoints on the profiles of these streams indicate that rejuvenation in the present erosion cycle is proceeding upstream toward the Hills. Active downcutting is taking place only in the downstream sections near the Belle Fourche and Cheyenne South Fork rivers. Because the latest rejuvenation is expressed by terracing of the Farmingdale surface in both major and minor streams of the area, it follows that this rejuvenation must be related to a wave of rejuvenation which has proceeded up the Cheyenne River but which has not yet reached the headwaters of its tributaries. A direct connection is therefore established between terrace cycles in the Missouri and Cheyenne River valleys. Todd (1923, pp. 491–493) has shown that the Missouri Valley terraces are genetically related to the Wisconsin stage of continental glaciation. Presence of the Wisconsin ice sheet in eastern South Dakota resulted in ponding of the Missouri River waters along its western border. Retreat of the ice permitted the Missouri River to reexcavate its valley, leaving behind terraces which mark the glacial stages. Fillman (1929, p. 44) states: "The possibility presents itself that the valley terraces of the Black Hills region are genetically related with the glacial terraces farther east and downstream." The presence of the knickpoints in the Black Hills streams makes this possibility much more probable. It is highly probable that the Farmingdale terrace is correlative with the last glacial stage in eastern South Dakota. The inference is strong that all the Pleistocene terraces in this area (Rapid, Sturgis, Bear Butte, and Farmingdale) may owe their origin to downstream blocking by Pleistocene ice sheets.

MacClintock (1936, pp. 346–360) suggests that differential uplift has occurred in the region since late Wisconsin time. As proof he cites the existence of varved sediments in the White River Valley. He believes that these sediments, which are found at several places from the White River headwaters to the Missouri River, were deposited in an ice-dammed lake resulting from a late Wisconsin glacial invasion. He discounts the idea that the varves could have been deposited in local lakes, and thus he postulates a rise of the western part of the basin with respect to the eastern part of 2,000 feet in post-Wisconsin time. A differential uplift of this type should be reflected in the Black

Hills region by renewed downcutting of the streams in their headward portions. However, this is not the case; post-Wisconsin downcutting in those streams is limited to the sections below the knickpoints. Therefore, post-Wisconsin uplift is not indicated in the Black Hills region.

### DRAINAGE PATTERNS

Most streams flowing eastward from the Black Hills are characterized by asymmetrical north-south profiles. The north-facing slopes of the valleys are steeper than the south-facing slopes.

Where asymmetry is greatest, the valleys are characterized by stream terraces which appear only on the north sides. Thus asymmetry is here the topographic expression of the unequal lateral distribution of stream terraces. Rapid Creek Valley (fig. 2) furnishes the best illustration of this phenomenon. The Rapid and Sturgis terraces are confined almost exclusively to the north side of the valley. An asymmetrial terrace distribution of this type must result from lateral southward shifting of the stream axis. Evidence supporting this hypothesis is shown in figure 4. As pointed out previously, the anomalous terraces on the south side of Rapid Creek are best explained by the hypothesis of a southward-shifting stream axis.

Asymmetry of valleys has been ascribed to many causes. Where lateral shifting of stream axis is evidenced, the possible causes may be reduced to the effects of regional dip, earth rotation, and regional tilting.

The regional dip in this area is east. Thus the effects of regional dip must be discounted because the lateral shift in stream axes is at right angles to the regional dip.

In the northern hemisphere streams are deflected to the right by the Coriolis force, an apparent force due to the earth's rotation. Thus, in an eastward-flowing stream, the steepest side of the valley should appear on the south, which is as observed in the streams under consideration. If this force has been effective in producing asymmetry, it should operate on all streams in the area, regardless of direction of flow. This is not the case, however, in the Black Hills region. Asymmetry is here limited to those streams flowing eastward from the Hills. Therefore, the Coriolis force must be ruled out as a major cause of asymmetry in this area.

The remaining possibility is regional downward tilt to the south. Unfortunately, no independent proof is available for such tilting, aside from the apparent southward shift of stream axes.

A corollary phenomenon of the asymmetrical valleys is the parallelism of tributary gulleys on the north sides of the valleys. As shown by the terrace outlines in Rapid Creek Valley (fig. 2) and Bear Butte Creek Valley (fig. 5), these gulleys trend north-northwest–south-southeast. A parallel stream pattern is usually controlled by structure, but this explanation does not apply here, as the formations are essentially flat-lying over the area.

W. L. Russell (1929, pp. 249–255) discussed the marked northwest-southeast alignment of valleys and ridges in the northwestern Great Plains. He found that this phenomenon was present in the minor drainage only. His explanation is that northwest-southeast aligned sand ridges were developed on a Tertiary surface by means of prevailing northwest winds and that the drainage pattern which originated in this way has so continued, although most of the Tertiary cover is now removed from most areas. Russell contends that this phenomenon cannot be due to tilting, for otherwise the

main streams would have shown the effects.

However, the asymmetry of the Black Hills valleys is best explained on the basis of regional tilt to the south, which probably took place after the main drainage lines of the present were established. On this basis parallel tributary gulleys are to be expected. It is postulated that, as the axes of the streams migrated southward, the tributary gulleys were lengthened by footward extension. This process would result in parallel extension of the minor drainage.

In summation, the channels of streams flowing eastward from the Black Hills have shifted southward, the migration probably starting in early Pleistocene time. The proof of this migration is found in the asymmetrical terrace distributions of the valleys and in certain anomalous terraces in Rapid Creek Valley. On the available evidence, regional downward tilt to the south is indicated.

## Sedimentology

### FIELD WORK

The field investigation involved the collection of samples for sedimentary analysis and the mapping of stream terraces. The latter problem was discussed in the physiographic section. The mapping of terraces was necessary for accurate sampling of the sediments. It was desirable to sample gravel from one terrace which could be easily traced over the area of study. The Sturgis terrace was selected for sampling. It is widespread along Bear Butte, Rapid, and Battle creeks and can be traced eastward from the Hills to the Belle Fourche and Cheyenne rivers.

Three types of samples were collected: (1) channel samples, (2) limestone pebble sets, and (3) spot sand samples.

Channel samples, weighing 150–200 pounds, were collected for mechanical and lithologic analysis. The sample portions greater than 2 mm. were hand-sieved and weighed in the field. The portions below 2 mm. were quartered for laboratory analysis, and the 16–32 mm. grades were saved for lithology counts.

Sets of Minnekahta (Permian) limestone pebbles were collected at most channel-sample localities and at additional localities for roundness and sphericity analysis. Fifty to one hundred pebbles of sizes 16–32 mm. and 32–64 mm. were collected at random from the exposures. The sphericity of each pebble was measured with an intercept gauge which measures three mutually perpendicular axes. As shown by Krumbein (1941*a*, pp. 64–72), ratios of these lengths may be used to compute the sphericity. Roundness was estimated visually by comparison with a standard set of pebble images of known roundness and sphericity.

A series of spot sand samples was collected from the present-day channel deposits of Battle Creek and the Cheyenne River. Samples of about 100 gm. (1–2 mm.) were taken from sand bars or banks.

### GRAVEL ANALYSIS

#### PROBLEMS OF SAMPLING

*Sample types.*—The choice of a sample type appropriate for mechanical analysis was limited by the nature of the deposits and their exposures. Over large areas the Sturgis gravels are cemented in varying degrees with calcium carbonate. Because of this consolidation it was impracticable from the standpoint of time and labor to take samples from locations other than gravel pits, road cuts, or steep terrace scarps cut by the present-day streams.

The channel samples collected in this

study reveal the average characteristics of the gravel deposits. It was impossible to deal with one sedimentation unit, which is defined as that thickness of sediment which was deposited under essentially constant physical conditions (Otto, 1938, pp. 569–582). Insufficient exposures were available to trace such a unit, and at some sample locations not all the gravel deposit was exposed.

In Rapid Creek an approach to the sedimentation unit was realized, because most of the exposures in the Sturgis terrace revealed rather homogeneous gravel (table 2). The Sturgis gravels of Bear Butte Creek are homogeneous or poorly bedded (table 3), whereas the gravels of Battle Creek are poorly to moderately bedded (table 4). Exposures of the Farmingdale terrace are moderately to well bedded (pl. 1).

Sampling from a river terrace avoids one of the main difficulties of sampling the sediment of a present-day stream. In the latter the size of the material varies widely, depending on its location with regard to the main channel. On the other hand, a terrace deposit contains material which was deposited under varying conditions and over a wide range of time. Thus a channel sample through a terrace gives data on the average size of the load deposited by the stream at that point.

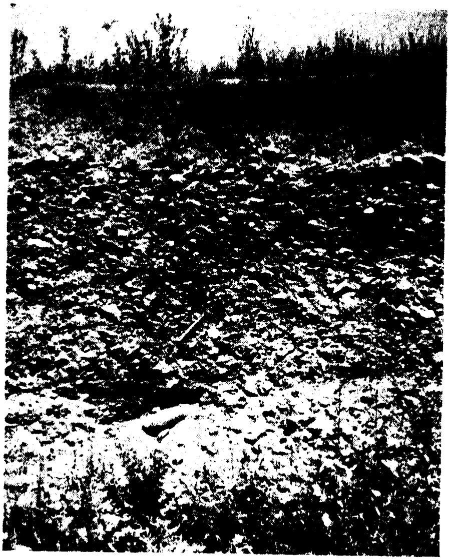
TABLE 2

CHANNEL-SAMPLE DESCRIPTIONS, RAPID CREEK

| Sample No. | Terrace | Thickness of Gravel Cap (Feet) | Length of Channel Sample (Feet) | Cementation | Bedding | Nature of Exposure |
|---|---|---|---|---|---|---|
| R-1......... | Sturgis | 11 | 8 | Medium | Homogeneous | Gravel pit |
| R-2......... | Sturgis | 25 | 5 near base; 3 at middle | Medium | Homogeneous | Terrace rim |
| R-3......... | Sturgis | 30 | 7 near top | Heavy | Poorly bedded; some 6-inch sand lenses | Railroad cut |
| R-4......... | Sturgis | 12 | 6 near top | Medium | Homogeneous | Road cut |
| R-5......... | Sturgis | 20 | 3, 3-foot channels, top to bottom | Light | Homogeneous | Road cut |
| R-6......... | Sturgis | 11 | 3, 2-foot channels, top to bottom | Light | Poorly bedded | Gravel pit |
| R-7......... | Farmingdale | 6 | 6 | None | Moderately bedded | Road cut |
| R-8......... | Farmingdale | 25+ | 7 near top | None | Well bedded | Road cut |
| R-9......... | Sturgis | 2 | 2 | Light | Homogeneous | Terrace rim |
| R-10........ | Farmingdale | 12 | 8 | None | Poorly bedded | Road cut |

*Sampling errors.*—Sampling errors fall into two categories: those due to the size of the sample and those due to local variations in the deposit.

The size of the sample for mechanical



A

B

C

D

PLEISTOCENE AND RECENT GRAVEL DEPOSITS

*A*, Sturgis terrace gravel, sample R-3, Rapid Creek; *B*, Farmingdale terrace gravel, sample R-9, Rapid Creek; *C*, flood plain gravel, sample BB-1, Bear Butte Creek; *D*, Sturgis terrace gravel, sample R-1, Rapid Creek.

analysis varied between 150 and 200 pounds, depending on the coarseness of the material. Wentworth's empirical rule was followed (Wentworth, 1926), which requires that a sample be large enough that several fragments of the largest grade are included. From probability considerations, this number of fragments should be at least four. It was not possible to include the largest boulders in the size-frequency distributions. To represent them correctly ( $>128$ mm.) would require a sample of several hundred pounds. However, because of their relative scarcity, it is assumed that the mean size of such a large sample would

TABLE 3

CHANNEL-SAMPLE DESCRIPTIONS, BEAR BUTTE CREEK

| Sample No. | Terrace | Thickness of Gravel Cap (Feet) | Length of Channel Sample (Feet) | Cementation | Bedding | Nature of Exposure |
|---|---|---|---|---|---|---|
| BB-1........ | Present-day flood plain | 20+ | 6 | None | Poorly bedded | Gravel pit |
| BB-2........ | Bear Butte | 10 | 8 | Heavy | Poorly bedded | House excavation |
| BB-3........ | Sturgis | 8 | 5 | Light | Homogeneous | Road cut |
| BB-4........ | Sturgis | 10+ | 6 | Light | Poorly bedded | Gravel pit |
| BB-5........ | Sturgis | 8 | 5 | Medium | Poorly bedded | Gravel pit |
| BB-6........ | Sturgis | 8 | 2, 3-foot channels | Heavy | Moderately bedded | Gravel pit |
| BB-7........ | Bear Butte | 11 | 8 | None | Well bedded | Terrace rim |

TABLE 4

CHANNEL-SAMPLE DESCRIPTIONS, BATTLE CREEK

| Sample No. | Terrace | Thickness of Gravel Cap (Feet) | Length of Channel Sample (Feet) | Cementation | Bedding | Nature of Exposure |
|---|---|---|---|---|---|---|
| B-1......... | Sturgis | 34+ | 10 near top; 4 near bottom | Light | Poorly bedded | Gravel pit |
| B-2......... | Sturgis | 12 | 5, 1-2-foot channels top to base | Light | Poorly bedded | Road cut |
| B-3......... | Sturgis | 20 | 4, 3-foot channels top to base | Light | Poorly bedded | Terrace rim |
| B-4......... | Sturgis | 20 | 4, 3-foot channels top to base | None | Moderately bedded | Terrace rim |
| B-5......... | Sturgis | 8 | 8 | None | Moderately bedded | Terrace rim |
| B-6......... | Sturgis | 12 | 5, 2-foot channels | None | Moderately bedded | Terrace rim |

differ little from that of a smaller sample (150–200 pounds) which did not include them. This assumption is supported by Krumbein's work on the flood deposits of Arroyo Seco (Krumbein, 1942, pp. 1355–1402).

To determine what effect local variation had on the median size, the probable error was computed for the deposits at sample site R-3 of Rapid Creek. The probable error measures the chance deviation of a given sample from the average value of the material being sampled. Three 6-foot channel samples, averaging 165 pounds each, were collected from the exposure. The channels were separated horizontally by 20 feet. Standard procedures for computing the probable error were used (Krumbein, 1934, pp. 204–214). The probable error of the median size was found to be less than 2 per cent. The validity of this small sampling error may be challenged statistically because this method of error determination holds strictly only for normal distributions. The Sturgis gravel deposits do not yield normal size-frequency curves but highly skewed distributions. However, it can at least be concluded that the sampling errors due to chance deviations are small.

The standard error of the mean was used to compute sampling errors of shape and roundness. It is defined as that error which will not be exceeded in 67 cases out of 100. The mean roundness computed from 50 pebbles had a standard error of less than 1 per cent. The mean sphericity had a standard error of less than 2 per cent. The standard errors of mean roundness and sphericity of quartz sand grains were about the same as those for pebbles.

By comparing these errors with the tabulated values of size, shape, and roundness, it is seen that the errors in sampling are small enough that no significant trends of these parameters with distance will be obscured.

## ANALYSIS AND PRESENTATION OF DATA

*Size.*—In the laboratory the quartered samples (< 2 mm.) were analyzed into Udden grades to complete the frequency distributions. The size-frequency distributions contain material in twelve size grades, from 128 mm. to < $\frac{1}{16}$ mm. (tables 5, 6, and 7).

The data show that two modes are present in most of the frequency distributions. The principal mode is in the largest sizes. A weak secondary mode appears in the sizes $\frac{1}{4}$–2 mm. In Rapid and Bear Butte creeks the secondary modes occur mainly in the $\frac{1}{4}$–1-mm. grades. In Battle Creek the secondary modes are primarily in the 1–2-mm. grades.

It was suggested by Udden (1914, pp. 736–737) that bimodal size-frequency distributions may represent a combination of traction and suspension loads. In the San Gabriel flood gravels, Krumbein (1940, pp. 639–676) found that most of the exposures contained pebbles completely surrounded by sand. Therefore, he thought it unlikely that the material represented two periods of deposition, such as openwork gravel with later infiltration of sand, and he concluded that an abrupt velocity decrease resulted in simultaneous deposition of both traction and suspension loads.

On the other hand, Fraser (1935, pp. 910–1010) contends that simultaneous deposition of large pebbles and fine sand is improbable, since the velocity of a stream carrying pebbles 10 inches in diameter would have to be decreased 60 per cent before 1-mm. size sand could be deposited. He considers it unlikely that such violent changes in current velocity always occur when coarse material is deposited. He concludes, therefore, that at

## TABLE 5

### Size Composition of Sturgis Terrace Deposits, Rapid Creek
(Expressed as Weight Percentages)

| Sample No. | Grade Limits in Mm. | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 128–64 | 64–32 | 32–16 | 16–8 | 8–4 | 4–2 | 2–1 | 1–½ | ½–¼ | ¼–⅛ | ⅛–1/16 | <1/16 |
| R-1 | 25.9 | 31.1 | 18.2 | 10.0 | 4.5 | 2.2 | 1.4 | 1.6 | 2.3 | 1.2 | 0.7 | 0.9 |
| R-2 | 27.0 | 25.5 | 14.7 | 8.6 | 5.7 | 3.9 | 2.8 | 3.7 | 3.9 | 1.9 | 1.0 | 1.3 |
| R-3 | 24.9 | 28.5 | 17.2 | 8.6 | 5.1 | 3.1 | 2.9 | 4.2 | 3.5 | 0.8 | 0.5 | 0.7 |
| R-4 | 21.4 | 33.1 | 14.6 | 7.8 | 5.0 | 3.6 | 3.4 | 5.0 | 3.9 | 1.1 | 0.6 | 0.5 |
| R-5 | 14.2 | 18.2 | 18.3 | 14.5 | 8.8 | 5.3 | 3.8 | 5.5 | 6.0 | 2.5 | 1.6 | 1.3 |
| R-6 | 17.6 | 15.7 | 19.8 | 12.3 | 8.5 | 6.5 | 5.3 | 6.2 | 5.2 | 1.7 | 0.7 | 0.5 |
| R-7* | 7.3 | 20.6 | 21.4 | 15.1 | 9.2 | 6.3 | 4.6 | 6.0 | 6.9 | 1.5 | 0.6 | 0.5 |
| R-8* | 5.6 | 18.6 | 19.8 | 16.0 | 11.7 | 7.3 | 4.5 | 5.1 | 6.2 | 3.2 | 1.3 | 0.7 |
| R-9 | 11.6 | 18.3 | 17.3 | 9.4 | 8.5 | 7.6 | 6.6 | 6.2 | 6.9 | 4.9 | 1.8 | 0.9 |
| R-10* | ...... | 11.7 | 21.8 | 19.1 | 11.6 | 6.4 | 4.7 | 9.0 | 10.2 | 3.3 | 1.4 | 0.8 |

* Farmingdale terrace.

## TABLE 6

### Size Composition of Sturgis Terrace Deposits, Bear Butte Creek
(Expressed as Weight Percentages)

| Sample No. | Grade Limits in Mm. | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 256–128 | 128–64 | 64–32 | 32–16 | 16–8 | 8–4 | 4–2 | 2–1 | 1–½ | ½–¼ | ¼–⅛ | ⅛–1/16 | <1/16 |
| BB-1* | 33.8 | 29.2 | 14.9 | 6.9 | 3.6 | 2.4 | 1.9 | 2.0 | 2.0 | 1.6 | 1.0 | 0.5 | 0.2 |
| BB-2 | ....... | 25.5 | 29.4 | 18.1 | 10.1 | 6.1 | 3.2 | 1.7 | 1.6 | 1.5 | 0.9 | 0.9 | 1.0 |
| BB-3 | ....... | 16.1 | 21.7 | 19.8 | 11.6 | 6.7 | 4.6 | 3.9 | 5.0 | 5.4 | 2.8 | 1.4 | 1.0 |
| BB-4 | ....... | 9.8 | 21.8 | 20.2 | 14.1 | 8.6 | 6.5 | 5.8 | 5.9 | 3.6 | 1.8 | 1.2 | 0.7 |
| BB-5 | ....... | 14.5 | 26.3 | 22.6 | 13.2 | 8.3 | 4.4 | 2.2 | 2.7 | 3.0 | 1.3 | 0.9 | 0.6 |
| BB-6 | ....... | 6.0 | 12.0 | 17.3 | 11.8 | 9.1 | 9.1 | 13.6 | 14.8 | 4.2 | 1.2 | 0.6 | 0.3 |
| BB-7† | ....... | 7.1 | 14.1 | 20.2 | 17.1 | 11.0 | 7.7 | 5.8 | 6.0 | 5.9 | 3.0 | 1.4 | 0.7 |

* Present-day stream gravel.

† Bear Butte terrace.

## TABLE 7

### Size Composition of Sturgis Terrace Deposits, Battle Creek
(Expressed as Weight Percentages)

| Sample No. | Grade Limits in Mm. | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 256–128 | 128–64 | 64–32 | 32–16 | 16–8 | 8–4 | 4–2 | 2–1 | 1–½ | ½–¼ | ¼–⅛ | ⅛–1/16 | <1/16 |
| B-1 | 28.3 | 25.4 | 11.5 | 7.5 | 5.9 | 4.4 | 4.5 | 4.7 | 3.5 | 2.3 | 1.1 | 0.6 | 0.3 |
| B-2 | ....... | 26.4 | 21.5 | 14.2 | 7.1 | 5.6 | 5.2 | 5.4 | 6.0 | 4.5 | 1.9 | 1.4 | 0.8 |
| B-3 | ....... | 12.0 | 14.9 | 12.1 | 10.6 | 10.4 | 10.9 | 12.2 | 8.9 | 4.4 | 1.8 | 1.1 | 0.7 |
| B-4 | ....... | 9.5 | 10.6 | 13.8 | 12.6 | 10.8 | 10.5 | 10.5 | 8.7 | 7.8 | 3.3 | 1.3 | 0.6 |
| B-5 | ....... | 7.2 | 9.2 | 16.6 | 18.6 | 14.3 | 11.7 | 8.9 | 6.4 | 3.0 | 1.4 | 1.5 | 1.2 |
| B-6 | ....... | 7.9 | 9.8 | 14.9 | 15.7 | 8.6 | 9.6 | 12.2 | 11.0 | 5.7 | 2.4 | 1.3 | 0.9 |

any given instant a river usually deposits only a very limited size range of material and that the finer sizes present in gravels are the result of later infiltration. He points out that the time of infiltration is a function of the current velocity and that, if the velocity fluctuates widely, the infilling of smaller sizes may follow closely after the deposition of the coarse material.

It is believed by the writer that the major portion of the fine sizes could not have been added to the deposit until the current velocity had decreased beyond their critical velocities for motion. Thus infiltration of successively smaller particles would rapidly follow deposition of the larger particles. By means of the Black Hills data one can determine whether or not the amount of material found in the secondary modes is of the right order of magnitude to represent infiltration of sand into gravel interstices.

The subject of porosity as related to systematic packing of spheres has been discussed by Graton and Fraser (1935, pp. 785–909). They found that, with systematic packing of equal-sized spheres, the loosest packing had a porosity of 47.64 per cent and the tightest packing a porosity of 25.95 per cent. If one assumes a mixture of two sizes of spherical particles having the same density—the large size making up the matrix and packed systematically and a much smaller size packed systematically within the interstices of the large size—then one can calculate the weight percentage per unit volume of each size fraction. For example, Case 1 of Graton and Fraser is selected. The unit cell of this cubic packing contains 8 spheres of radius $R$. The volume of the unit cell is $8.00R^3$, the volume of the unit void is $3.81R^3$, and the porosity is 47.64 per cent. Consider the unit void to be packed with small spheres of density equal to that of the large spheres and assume that the small spheres also have a cubic packing. If the size of the large spheres is very large compared to the size of the small spheres, then haphazard packing near the walls of the "container" will be relatively unimportant. Then 53 per cent of the unit volume will be occupied by large spheres. Of the remaining 47 per cent, 53 per cent will be occupied by small spheres and 47 per cent will be void. Thus only 25 per cent (i.e., 53 per cent *times* 47 per cent) of the unit volume will be occupied by small spheres. Since the densities of the spheres are assumed to be equal, then

$$\text{Weight ratio} = \frac{\text{Per cent small spheres}}{\text{Per cent small spheres} + \text{per cent large spheres}}$$
$$= \frac{0.25}{0.25 + 0.53} = 32 \text{ per cent}\,.$$

Thus for the loosest type of systematic packing the small spheres constitute 32 per cent by weight of the total weight of the mixture.

For the tightest systematic packing of spheres (Case 6: Rhombohedral), an analogous calculation yields a weight ratio of 22 per cent.

It is obvious that the simplified conditions of the above analysis cannot be applied directly to a natural gravel. In a natural gravel there exists an almost infinite number of sizes, the particles are not spheres, the densities are different, and the packing may vary from systematic to chaotic. In spite of these complexities, an interesting comparison can be made of the theoretical and actual situations.

If one considers the median size as representing an average matrix size and the secondary mode as representing material infiltrated into the matrix, the agreement between theoretical and actual weight percentages is remarkably close. In the natural gravels the fraction assumed as matrix-void filler includes material from the "low" class to the finest-size class (i.e., Rapid Creek, sample R-5: 2 mm. to < $\frac{1}{16}$ mm.; weight per cent, 20.7). The average weight percentage of analogous size classes from 16 samples in Rapid and Bear Butte creeks is 20 per cent. This weight ratio is quite comparable to the theoretical ratios of 22 and 32 per cent. A lower actual weight ratio is to be expected, since, when more than one size is involved, the porosity is generally decreased. In general, the weight ratios of the infilling fractions increase with decreasing mean size of the sediment. Although theoretically the porosity is independent of the grain size, it usually increases with decreasing grain size because of frictional and bridging effects.

In Rapid and Bear Butte creeks, thirteen samples out of sixteen have a "low" class at 1–2 mm. Although this low class may emphasize the secondary mode, it is not the primary cause of the latter. The lack of material in the 1–2 mm. class is probably due to the size character of grain aggregates and individual grains. Primary aggregates of grains, as in igneous rocks, are usually composed of grains less than 1 mm. in size. When the aggregates, which are usually larger than 2 mm., break up, only particles less than 1 mm. result.

As shown in table 7, the size composition of Battle Creek is quite unlike that of Rapid and Bear Butte creeks. The reason for this dissimilarity is twofold. First, Battle Creek and its tributaries, unlike the other creeks, drain a large area of granite. The granite generally has a coarse texture and is in part pegmatitic. Because of its coarse texture the granite furnishes abundant material to the 1–2-mm. size class, and therefore the persistent 1–2-mm. "low" class is missing in the Battle Creek size analyses. Second, the gravel deposits of Battle Creek are better bedded than are those of the other creeks. Layers of coarse sand commonly appear in the gravel exposures. Thus the size distributions of Battle Creek may represent a mixture of two types of sedimentation units, the gravel layers contributing to the coarse-size modes and the sand layers making up the secondary modes.

The size data have been analyzed statistically in terms of the moment measures $M_\phi$ ("phi mean"), $\sigma_\phi$ ("phi standard deviation"), and $Sk_\phi$ ("phi skewness") after Krumbein and Pettijohn (1938, pp. 242–252). The statistical parameters are summarized in table 8 and plotted as functions of distance in figures 9, 11, and 12.

In all cases the mean size decreases with distance from the Black Hills (fig. 9). The marked increase in the mean size from sample 4 to sample 5 in Bear Butte Creek is due to the addition of coarse gravel from the slopes of Bear Butte.

It has been shown by other investigations that in some present-day streams the mean weight of sediment decreases exponentially as a function of distance, according to Sternberg's law:

$$W = W_0 e^{-kx} ,$$

where $W$ = weight at any point, $W_o$ = initial weight, $x$ = distance, and $k$ = coefficient of weight reduction. The exponential nature of this function requires that the data plot as a straight line on

semilog paper, where distance is the independent variable.

Barrell (1925, pp. 279–342) attributed this exponential size decrease to abrasion. More recently, the importance of selective transport has been recognized (Krumbein, 1937, pp. 577–601). The effects of abrasion alone on size reduction have been studied in the laboratory. In tumbling-barrel studies, where the selective action is absent, Krumbein (1941*b*, pp. 482–520) and Sarmiento (1945, pp. 1–59) found that size reduction of pebbles proceeded exponentially. A similar exponential size decrease was noted by Pettijohn and Lundahl (1943, pp. 69–78) in a study of Lake Erie beach sands. In this case, where the size reduction was large compared to the distance traveled, it is probable that selective transport was the main factor.

TABLE 8

STATISTICAL PARAMETERS OF SIZE DATA

| Sample No. | Phi Mean ($M_\phi$) | Phi Standard Deviation ($\sigma_\phi$) | Phi Skewness ($Sk_\phi$) | Geometric Mean Diameter in Mm. ($GM_\xi$) |
|---|---|---|---|---|
| | Rapid Creek | | | |
| R-1.... | −4.58 | 2.26 | 0.939 | 23.91 |
| R-2.... | −4.12 | 2.69 | 0.687 | 17.38 |
| R-3.... | −4.31 | 2.42 | 0.625 | 19.83 |
| R-4.... | −4.18 | 2.32 | 0.833 | 18.12 |
| R-5.... | −3.28 | 2.78 | 0.506 | 9.71 |
| R-6.... | −3.48 | 2.78 | 0.332 | 11.16 |
| R-7.... | −3.23 | 2.49 | 0.431 | 9.38 |
| R-8.... | −2.99 | 2.56 | 0.429 | 7.94 |
| R-9.... | −2.87 | 2.82 | 0.341 | 7.31 |
| R-10... | −2.34 | 2.56 | 0.310 | 5.06 |
| | Bear Butte Creek | | | |
| BB-1... | −5.70 | 2.36 | 0.977 | 51.96 |
| BB-2... | −4.52 | 2.24 | 0.924 | 22.94 |
| BB-3... | −3.52 | 2.76 | 0.515 | 11.47 |
| BB-4... | −3.38 | 2.52 | 0.472 | 10.41 |
| BB-5... | −4.00 | 2.29 | 0.720 | 16.00 |
| BB-6... | −2.47 | 2.81 | 0.327 | 5.54 |
| BB-7... | −1.89 | 2.57 | 0.376 | 3.71 |
| | Battle Creek | | | |
| B-1.... | −5.04 | 2.76 | 0.611 | 32.89 |
| B-2.... | −3.80 | 2.83 | 0.499 | 11.06 |
| B-3.... | −2.77 | 2.65 | 0.175 | 6.82 |
| B-4.... | −2.43 | 2.70 | 0.140 | 5.39 |
| B-5.... | −2.52 | 2.48 | 0.189 | 5.74 |
| B-6.... | −2.41 | 2.64 | 0.103 | 5.31 |

It is evident from the graphs of $M_\phi$ against distance (fig. 9) that in these streams size reduction does not follow Sternberg's law. In each case the upstream portion of the curve is steeper than the downstream portion. Sternberg stated that the gradient of a graded river also follows a logarithmic curve. On this basis, Barrell (1925, pp. 279–342) concluded that both size of gravel and gradient of a single stream are related under the same general law. Shulits (1941, pp. 622–630) discussed this relation and obtained an equation of the slope of river profiles:

$$S = S_0 e^{ax} ,$$

where $S$ = slope at any point, $S_0$ = slope at some initial point upstream, $x$ = distance measured upstream, and $a$ = coefficient of slope change.

The slopes of the Black Hills streams, however, do not follow this exponential law. Nevertheless, since both mean size and slope decrease downstream, it is possible to describe one as a function of the other. Therefore, in figure 10, mean size ($M_\phi$) is plotted against gradient ($G$, table 9) as the independent variable. As a first approximation, straight lines can be drawn through these data, indicating that the relation is linear. The equation of each curve is of the form

$$M_\phi = a + bG ,$$

where $M_\phi$ = phi mean size, $G$ = gradient in feet per mile, $a$ = constant, and

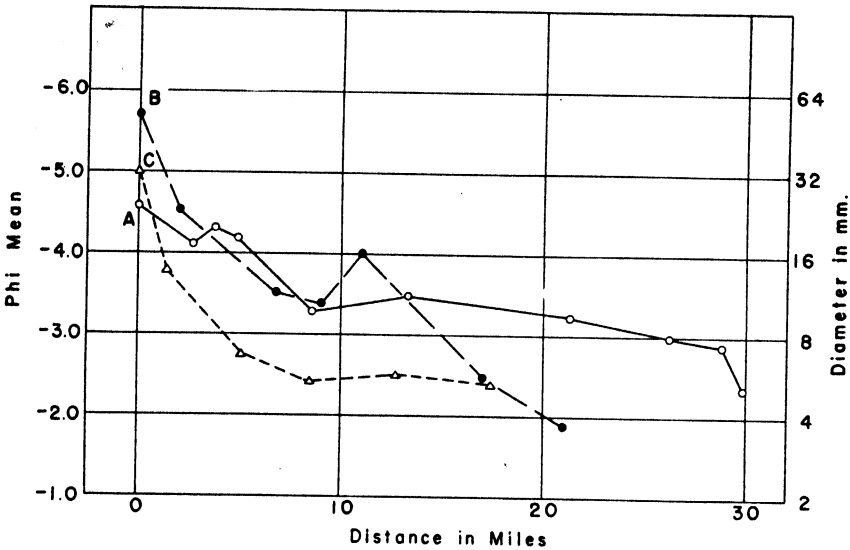



FIG. 9.—Relation of mean size to distance of transport. *A*, Rapid Creek; *B*, Bear Butte Creek; *C*, Battle Creek.

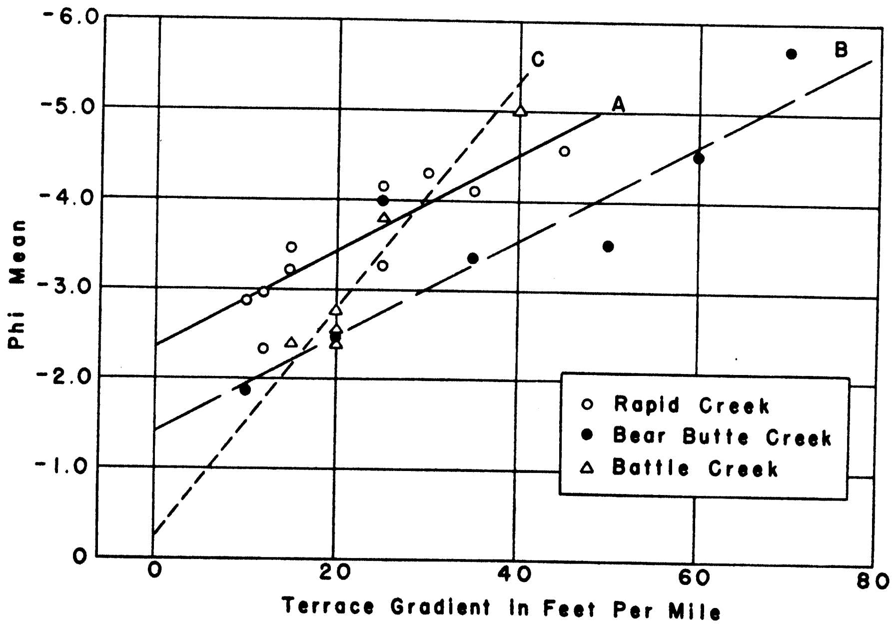



FIG. 10.—Relation of mean size to terrace gradient. *A*, Rapid Creek; B, Bear Butte Creek; *C*, Battle Creek.

$b$ = constant. The constant $b$ is the slope of the curve. The constant $a$ corrects for the curve not passing through the origin and has the units of $M_\phi$. The value of $a$ is read graphically where the curve intersects the line $G = 0$. Computed values of $a$ and $b$ are listed in table 10.

Although the curves of Rapid and Bear Butte creeks (fig. 10) have the same slope, they are not identical. For a specific value of $G$ the mean size of Rapid Creek gravel is greater than that of Bear Butte Creek. This is due to the fact that Rapid Creek has a greater mean discharge than has Bear Butte Creek. Thus the decrease in mean gravel size is related to three factors: (1) abrasion, (2) stream gradient, and (3) mean discharge.

TABLE 9

TERRACE GRADIENTS OF STREAMS IN THE BLACK HILLS REGION

| Sample No. | Terrace Gradient (Feet/Mile) $G$ |
|---|---|
| *Rapid Creek:* | |
| R-1 | 45 |
| R-2 | 35 |
| R-3 | 30 |
| R-4 | 25 |
| R-5 | 25 |
| R-6 | 15 |
| R-7 | 14 |
| R-8 | 12 |
| R-9 | 10 |
| R-10 | 12 |
| *Bear Butte Creek:* | |
| BB-1 | 70 |
| BB-2 | 60 |
| BB-3 | 50 |
| BB-4 | 35 |
| BB-5 | 25 |
| BB-6 | 20 |
| BB-7 | 10 |
| *Battle Creek:* | |
| B-1 | 40 |
| B-2 | 25 |
| B-3 | 20 |
| B-4 | 20 |
| B-5 | 20 |
| B-6 | 15 |

It can be inferred from the observed relation between mean size and stream gradient that Sternberg's exponential law of size decrease holds only in so far as the stream gradient also decreases exponentially. Therefore, such exponential relationships can be considered only as special cases of a more general law relating size and stream gradient.

The Phi Standard Deviations ($\sigma_\phi$) show no significant differences in the three creeks (table 8). The average $\sigma_\phi$ for all samples is about 2.6, which shows that the main part of the frequency distribution is included within 5.2 Udden grades. Thus the gravel is poorly sorted. In figure 11 the $\sigma_\phi$ is plotted against distance. The values of $\sigma_\phi$ are quite variable and show no systematic changes as a function of distance.

TABLE 10

NUMERICAL CONSTANTS OF THE MEAN SIZE, TERRACE GRADIENT EQUATION

| | $a$ | $b$ |
|---|---|---|
| Rapid Creek | −2.35 | 0.533 |
| Bear Butte Creek | −1.40 | 0.533 |
| Battle Creek | −0.25 | 1.25 |

All the samples have a positive skewness (table 8), which means that the size distributions are asymmetrical in the direction of the largest sizes. When $Sk_\phi$ is plotted against distance (fig. 12), it is observed that it declines rapidly as the material is transported away from the Hills. If an originally skewed distribution at the source can be assumed, this decrease in $Sk_\phi$ can be explained as a combined effect of abrasion and selective transportation. It has been shown experimentally by Sarmiento (1945, pp. 1–59) that the largest sizes of material are abraded the most rapidly. Thus it could be expected that by abrasion alone, the larger particles would decrease in size

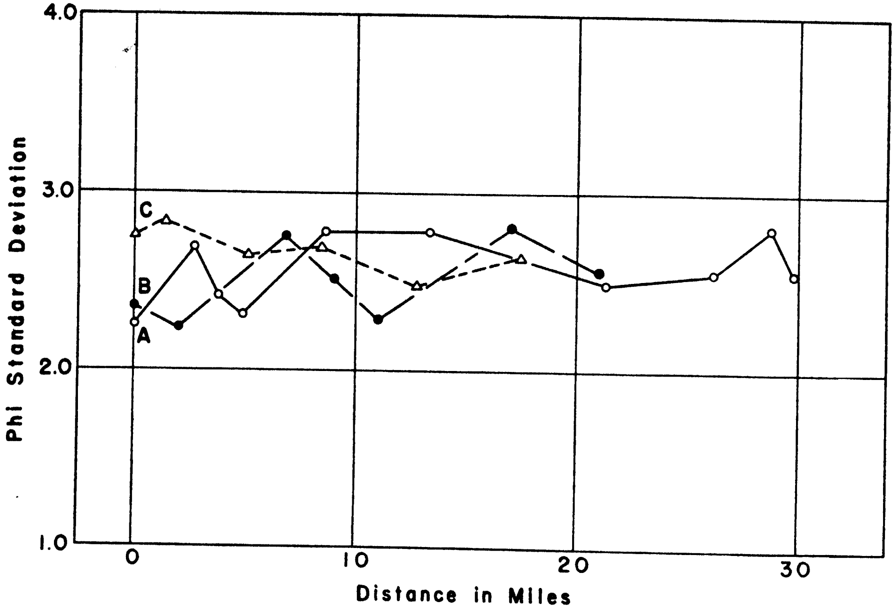



FIG. 11.—Relation of sorting to distance of transport. *A*, Rapid Creek; *B*, Bear Butte Creek; *C*, Battle Creek.

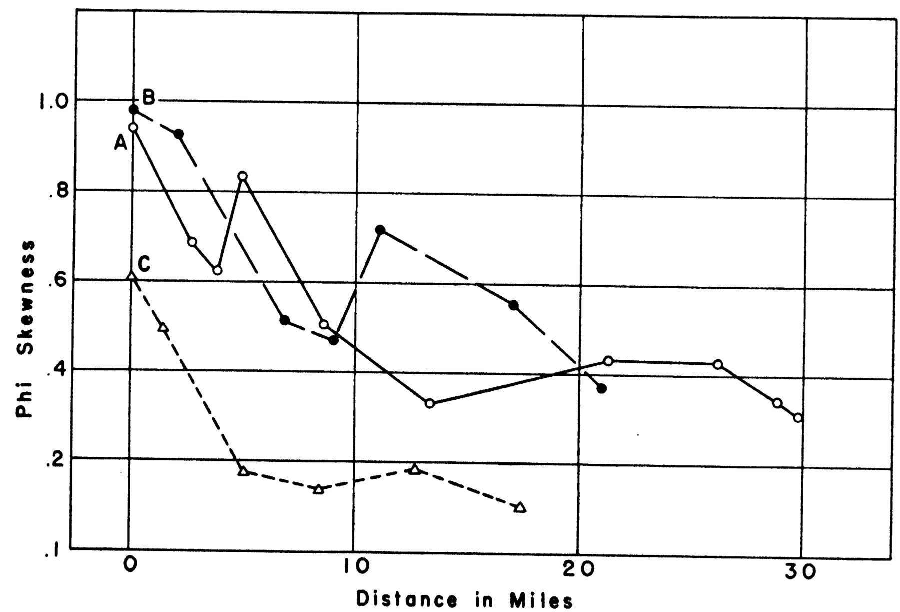



FIG. 12.—Relation of skewness to distance of transport. *A*, Rapid Creek; *B*, Bear Butte Creek; *C*, Battle Creek.

more rapidly than would smaller particles, thus tending to normalize the size distribution. Probably more important than wear is the effect of selective transport. In stream transport the smaller particles outrun the larger ones, which also works to normalize the size distribution.

Although no size distributions were obtained at the ultimate sources of the crushed material follows a geometric law in its size classes. Each size grade contains about half as much material by weight as does the next larger grade. The skewness data from the Black Hills gravel strongly suggest that the talus slopes furnishing material to the streams have such a skewed size distribution. The action of the stream in transporting this material is to reduce the skewness of

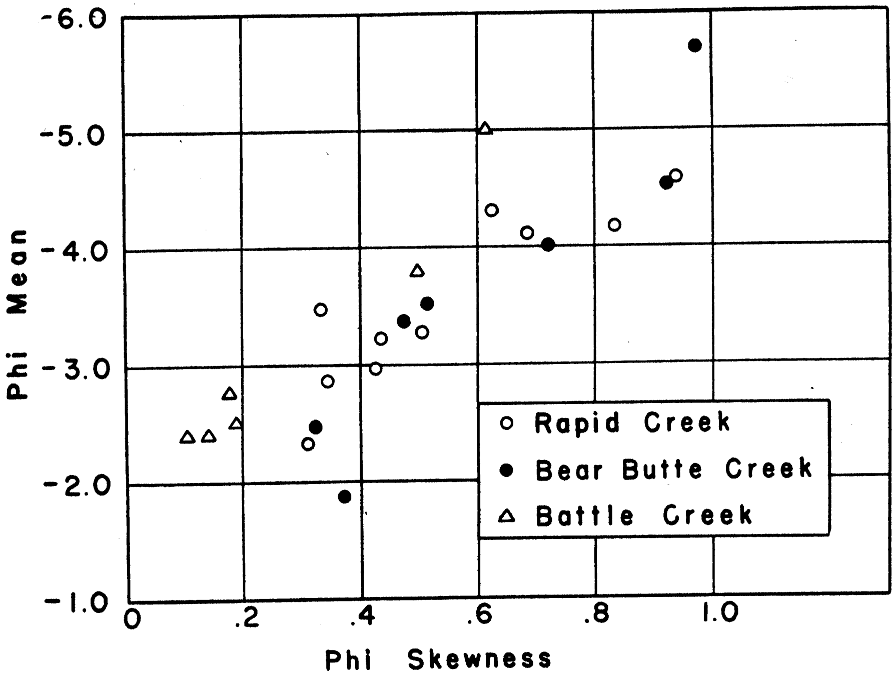



FIG. 13.—Relation of mean size and skewness (Black Hills data)

terrace gravels, some light is thrown on the problem by Krumbein and Tisdel (1940, pp. 296–305). Their experiments show that, when certain types of material (coal and quartz) are randomly crushed, the resulting size distributions are quite similar to those found in the gravel near the Black Hills. The coal and quartz broke up into fragments according to Rosin's law, which states that the distribution by differential wear and selective transport.

It is apparent from the data that skewness is a function of the mean size. In figure 13, $M_\phi$ is plotted against $Sk_\phi$ in a scatter diagram. It is seen that the samples having the largest mean size also have the most highly skewed distributions.

Additional information on the relation

of skewness to size can be obtained from a study of gravels at one location. To demonstrate this, use was made of data obtained by Kurk (1941, pp. 1–37) from Pleistocene outwash gravels at Cary, Illinois. Kurk sampled a series of sedimentation units from a vertical channel through the deposit. The writer made use of the mechanical analyses from these units to compute the phi skewness. Most of the individual sedimentation units had a bimodal size distribution. The skewness calculated from such a bimodal distribution is not representative of the skewness

TABLE 11

MEAN SIZE AND SKEWNESS DATA (AFTER KURK)

| Phi Mean ($M_\phi$) | Phi Skewness ($Sk_\phi$) |
|---|---|
| −5.60 | +0.84 |
| −4.46 | +0.55 |
| −4.32 | +0.93 |
| −3.41 | +0.54 |
| −3.06 | +0.57 |
| −2.94 | +0.26 |
| −2.01 | +0.34 |
| −1.95 | +0.41 |
| −1.59 | +0.05 |
| −0.27 | +0.03 |
| +0.70 | −0.18 |
| +2.47 | −0.01 |

of either of the modes exhibited in the unit but is rather a measure of the relative importance of the two modes. If the mode in the large size is the primary one, the phi skewness will be positive.

The $M_\phi$ and $Sk_\phi$ values calculated for these individual sedimentation units are listed in table 11. These data are plotted as a scatter diagram in figure 14.

It is apparent that a decrease in mean size is followed by a corresponding decrease in skewness. Thus it is evident that the decrease in the skewness of the Black Hills samples with distance is a function of the decrease in mean size.

*Lithology.*—The size fraction (16–32 mm.) of each channel sample was analyzed for its lithologic content. The results of this analysis are given in tables 12, 13, and 14. Two hundred pebbles per sample were identified from the gravels of Bear Butte Creek, four hundred from Rapid Creek, and six hundred from Battle Creek. If we overlook the changes in lithology with distance of transport, the data reveal pronounced differences in the lithologies of the various creeks. This is a direct result of unlike source rocks in the areas which the creeks drain.

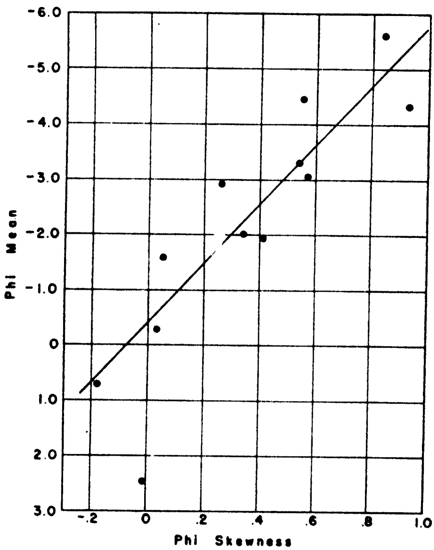



FIG. 14.—Relation of mean size and skewness. Pleistocene outwash gravel, Cary, Illinois (Kurk).

The samples nearest the Black Hills of each creek show most significant differences. In sample 1 of Bear Butte Creek, 54 per cent of the pebbles are limestone and 27 porphyry. The high percentage of limestone in this sample is due to the fact that limestone formations, which are exposed at only one place in hogbacks farther south, are repeated by anticlinal and synclinal structures in the drainage area of Bear Butte Creek. Thus more limestone is furnished to that stream.

Tertiary intrusive bodies in the northern Hills furnish the porphyry found in the gravels of Bear Butte Creek terraces.

Rapid Creek gravels are characterized by large percentages of quartzite, quartz, and pre-Cambrian metamorphic rocks (sample 1). Rapid Creek drains the central part of the Black Hills, which is predominantly composed of pre-Cambrian metamorphic rocks. Most of the quartzite tabulated here is of pre-Cambrian age.

Battle Creek drains the southern portion of the Hills. In sample 1 quartzite and quartz are conspicuous because large areas of the pre-Cambrian zone are with-

TABLE 12

BEAR BUTTE CREEK LITHOLOGY, STURGIS TERRACE
(Expressed as Number Per Cent)

| | Sample Number | | | | | | |
|---|---|---|---|---|---|---|---|
| | BB-1* | BB-2† | BB-3 | BB-4 | BB-5 | BB-6 | BB-7† |
| Pre-Cambrian metamorphics | 0.5 | 0.5 | 1.0 | 1.5 | ........ | ........ | 0.5 |
| Quartzite | 2.5 | 3.0 | 3.5 | 5.0 | 7.0 | 8.0 | 5.0 |
| Quartz | 3.0 | 2.0 | 5.5 | 2.0 | 5.5 | 5.0 | 6.0 |
| Porphyry and rhyolite | 27.0 | 36.5 | 46.0 | 58.5 | 48.5 | 26.0 | 47.0 |
| Limestone | 54.0 | 38.0 | 16.0 | 22.0 | 23.0 | 30.0 | 18.0 |
| Sandstone | 10.5 | 15.5 | 22.0 | 6.5 | 8.5 | 3.0 | 4.5 |
| Chert | 2.5 | 4.5 | 6.0 | 4.5 | 7.5 | 28.5 | 16.0 |
| Clay-ironstone | ........ | ........ | ........ | ........ | ........ | 2.0 | ........ |
| Chalcedony | ........ | ........ | ........ | ........ | ........ | 0.5 | ........ |
| Quartz-hematite | ........ | ........ | ........ | ........ | ........ | 2.0 | 3.0 |

* Present-day flood plain. † Bear Butte terrace.

TABLE 13

RAPID CREEK LITHOLOGY, STURGIS TERRACE
(Expressed as Number Per Cent)

| | Sample Number | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | R-1 | R-2 | R-3 | R-4 | R-5 | R-6 | R-7* | R-8* | R-9 | R-10* |
| Pre-Cambrian metamorphics | 34.0 | 37.5 | 31.7 | 34.0 | 16.8 | 16.0 | 11.2 | 8.8 | 7.0 | 7.0 |
| Quartzite | 22.5 | 20.5 | 33.7 | 34.3 | 49.0 | 50.5 | 42.5 | 34.3 | 28.5 | 29.7 |
| Quartz | 14.8 | 16.7 | 8.3 | 15.0 | 22.0 | 18.5 | 28.0 | 25.0 | 21.5 | 24.2 |
| Limestone | 17.7 | 14.0 | 11.3 | 3.7 | 3.0 | 3.7 | 3.0 | 5.0 | 0.5 | 1.3 |
| Sandstone | 8.5 | 5.7 | 11.7 | 8.5 | 4.7 | 3.7 | 0.8 | 1.0 | ...... | 1.3 |
| Chert | 2.5 | 4.3 | 3.3 | 4.5 | 4.5 | 5.3 | 12.7 | 20.5 | 15.8 | 23.5 |
| Clay-ironstone | ...... | 1.3 | ...... | ...... | ...... | 2.0 | 0.5 | 0.3 | 1.0 | 0.3 |
| Concretionary limestone (from Pierre shale) | ...... | ...... | ...... | ...... | ...... | 0.3 | 0.5 | 3.3 | 10.7 | 7.5 |
| Chalcedony | ...... | ...... | ...... | ...... | ...... | ...... | 0.5 | 1.8 | 13.8 | 4.2 |
| Porphyry | ...... | ...... | ...... | ...... | ...... | ...... | 0.3 | ...... | ...... | ...... |
| Feldspar | ...... | ...... | ...... | ...... | ...... | ...... | ...... | ...... | 1.2 | 1.0 |

* Farmindgale terrace.

in the Battle Creek drainage area. Also within this drainage area are large masses of pegmatitic Harney Peak granite. Because of the coarseness of this granite, the pebbles derived from it are either quartz or feldspar, with a few pebbles of graphic granite.

Farther out from the Hills the gravels include a few minor constituents from the Cretaceous Pierre shale and the Tertiary formations. These minor constituents are clay-ironstone, concretionary limestone, and chalcedony.

apparent increase in the amount of chert is due, therefore, to the loss of other rocks by abrasion and breakage. By comparing other rock types to chert as a standard, the relative resistances to abrasion and breakage of the former may be calculated. These calculations are shown graphically in figures 15–17. The chert ratio is plotted against distance as the independent variable. The chert ratio is calculated as

$$\text{Chert ratio} = \frac{X}{X+Y} \cdot 100 \,,$$

TABLE 14

BATTLE CREEK LITHOLOGY, STURGIS TERRACE
(Expressed as Number Per Cent)

| | Sample Numbers | | | | | |
|---|---|---|---|---|---|---|
| | B-1 | B-2 | B-3 | B-4 | B-5 | B-6 |
| Pre-Cambrian metamorphics | 4.0 | 2.7 | 5.7 | 8.8 | 4.0 | 5.2 |
| Quartzite | 22.3 | 17.3 | 23.8 | 34.5 | 38.5 | 42.2 |
| Quartz | 13.0 | 11.0 | 14.3 | 13.2 | 17.3 | 19.7 |
| Feldspar and graphic granite | 12.0 | 16.0 | 11.0 | 10.0 | 11.4 | 7.0 |
| Limestone | 23.3 | 22.2 | 21.0 | 14.0 | 11.3 | 9.0 |
| Sandstone | 23.5 | 28.3 | 20.5 | 11.7 | 10.5 | 6.8 |
| Chert | 1.7 | 2.5 | 3.7 | 7.3 | 6.7 | 9.5 |
| Clay-ironstone | ........ | ........ | ........ | 0.5 | 0.3 | 0.3 |
| Concretionary limestone | ........ | ........ | ........ | ........ | ........ | 0.3 |
| Tourmaline | 0.2 | ........ | ........ | ........ | ........ | ........ |

To determine the effects of transportation on gravel lithology and to obtain the relative resistances of various rock types to abrasion and breakage, ratios of one rock type to another were computed. The use of a ratio rather than a direct percentage avoids the effect of an apparent decrease of one rock type, which may be due only to the addition of new material of another rock type as the stream crosses an exposure of that new material. Chert was selected as a standard of reference because it is present in all the streams and is the hardest of all rocks present. An

where $X$ = the number of chert pebbles and $Y$ = the number of pebbles of some other rock type. The ratio is 100 if a certain rock type does not appear in a particular sample.

In Rapid Creek (fig. 15) the ascent of all the curves shows that the proportion of chert in the terrace deposits is increasing with distance from the Hills at the expense of all other rock types. It is, as anticipated, the most resistant to abrasion and breakage. The irregularities in the curve of any one rock type are due partly to local sample variation, partly

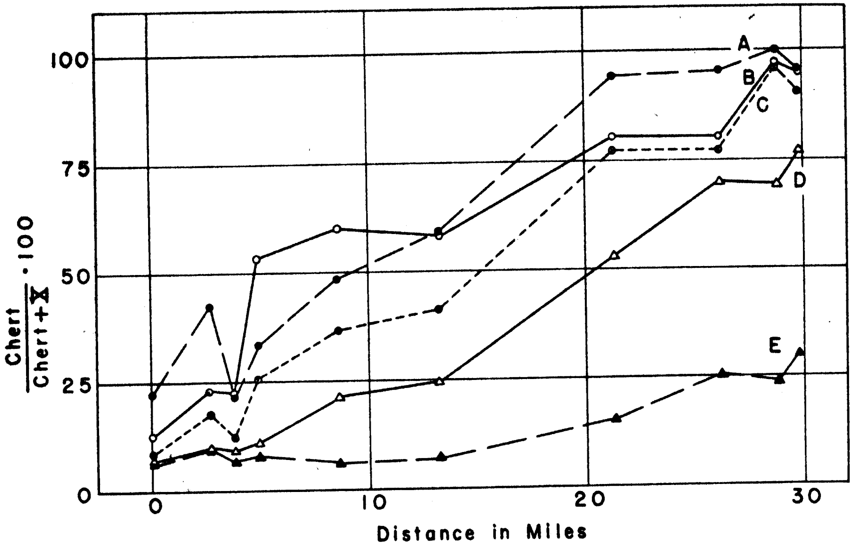



FIG. 15.—Effect of transport on rock constituents of gravel (Rapid Creek). Ratio of chert to chert plus each component: *A*, sandstone; *B*, limestone; *C*, sandstone plus limestone; *D*, pre-Cambrian metamorphics; *E*, quartz plus quartzite.

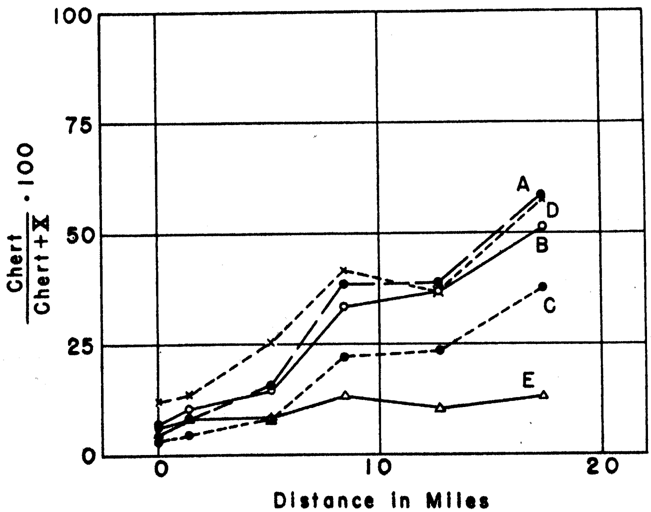



FIG. 16.—Effect of transport on rock constituents of gravel (Battle Creek). Ratio of chert to chert plus each component: *A*, sandstone; *B*, limestone; *C*, sandstone plus limestone; *D*, feldspar; *E*, quartz plus quartzite.

to influx of new material, and partly to the effects of increase or decrease of other types, which is not completely eliminated by the use of a ratio. This is illustrated by sample 3, which is from a locality about 1 mile east of the Dakota sandstone (Cretaceous) hogback. The sandstone ratio drops markedly from sample 2 to sample 3 because of the addition of sandstone to the stream from the Dakota outcrop. That the other ratios are affected by this addition is shown in the values in sample 3, where all ratios show a sympathetic decrease. East of sample locality 4, addition of limestone from the Colorado group (Cretaceous) may account for depression of the limestone curve. The sandstone curve reaches the 100 value at sample 9, indicating an absence of sandstone. This, however, does not mean that no sandstone pebbles (16–32 mm.) will be found in the gravel beyond this distance from the Hills. If a larger sample had been taken, some sandstone would undoubtedly have been found, but its percentage in the total sample would have been much reduced. It is to be expected that these curves will approach the 100 ratio asymptotically if large enough samples are taken. In general, quartz and quartzite are the most resistant to abrasion and breakage, followed by pre-Cambrian metamorphics, limestone, and sandstone. Although sandstone is composed mainly of hard quartz grains, it is the least resistant of all the rock types to abrasion and breakage, probably on account of its high friability.

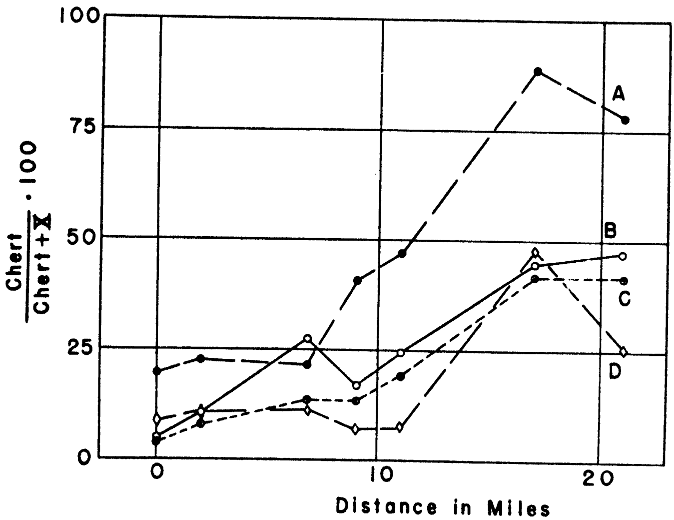



FIG. 17.—Effect of transport on rock constituents of gravel (Bear Butte Creek). Ratio of chert to chert plus each component: *A*, sandstone; *B*, limestone; *C*, sandstone plus limestone; *D*, porphyry.

The graph of Battle Creek is shown in figure 16. Sample 1 is from a locality east of the Dakota hogback, and therefore contamination from that source is absent. In general, the curves are similar to those of Rapid Creek, but the decrease in sandstone and limestone is less abrupt. This difference may be explained in two ways: Rapid Creek is a larger stream, and its gravel may have been subjected

to more vigorous transport. Moreover, the percentage of hard rock (quartzite, quartz, chert, pre-Cambrian metamorphics) in Rapid Creek is about 75 per cent near the Hills, while the same rocks in Battle Creek make up only about 40 per cent of the gravel. The limestone and sandstone pebbles of Rapid Creek are associated, therefore, with about twice as many hard-rock pebbles, which act as grinders to reduce them.

TABLE 15

STATISTICAL PARAMETERS OF SHAPE AND ROUNDNESS DATA, MINNEKAHTA LIMESTONE, RAPID CREEK

| Sample No. | 16–32 Mm. | | 32–64 Mm. | |
|---|---|---|---|---|
| | Mean Roundness ($P$) | Mean Sphericity ($\psi$) | Mean Roundness ($P$) | Mean Sphericity ($\psi$) |
| R-A... | 0.15 | 0.62 | 0.15 | 0.62 |
| R-B... | 0.34 | 0.59 | 0.37 | 0.57 |
| R-1.... | 0.39 | 0.51 | 0.49 | 0.56 |
| R-C... | 0.47 | 0.59 | 0.52 | 0.65 |
| R-D... | 0.46 | 0.53 | 0.48 | 0.59 |
| R-2.... | 0.52 | 0.58 | 0.57 | 0.62 |
| R-3.... | 0.55 | 0.62 | 0.59 | 0.65 |
| R-6.... | 0.62 | 0.61 | 0.63 | 0.66 |
| R-7.... | 0.64 | 0.61 | ........ | ........ |
| R-E... | 0.65 | 0.66 | ........ | ........ |
| R-8.... | 0.65 | 0.64 | ........ | ........ |
| R-10... | 0.67 | 0.63 | ........ | ........ |

Feldspar and graphic granite are an important component of Battle Creek gravels. Feldspar is seen to be about as resistant as limestone. Sandstone, as found in Rapid Creek, is slightly less resistant than limestone.

The graph of Bear Butte Creek is shown in figure 17. The curves in this creek are quite erratic, owing principally to contamination by new sources of material. The Dakota hogback lies between the sources of samples 2 and 3. The sandstone curve is depressed in this zone by the addition of sandstone from the Dakota formation. The depression of the limestone curve at sample 4 is due to additions from the Greenhorn limestone (Cretaceous). Bear Butte, a Tertiary intrusive of porphyry, is located about 3 miles north of sample 3. Contributions from this body account for the depression of the porphyry curve between sample localities 3 and 5. The incongruency of the ratios in sample 7 may be due to the fact that this deposit may consist of gravels from the Belle Fourche River rather than from Bear Butte Creek.

*Shape and roundness*.—The effects of transportation on the shape and roundness were studied for two sizes (16–32 and 32–64 mm.) of one type of rock. The type selected had to be fairly common and easily recognized, in order to make its collection possible. It was also necessary to have a source in the area where the pebbles were angular and to have only one source to avoid obscuring the results by the addition of fresh material at any other point. The Minnekahta limestone (Permian) satisfied these requirements. It outcrops as a continuous hogback around the Black Hills, averaging 40 feet in thickness. It has a distinctive gray color tinged with pink and purple, resulting in its being called the "purple limestone." Thin shale and fine sand partings give the limestone a tabular structure.

The statistical parameters of shape and roundness obtained by the study of samples from five streams are given in tables 15–17. In figures 18–20 these parameters are plotted as functions of distance of transport. The data from Rapid and Battle Creek samples are plotted separately, whereas those of Bear Butte Creek, the Belle Fourche River, and the Cheyenne River are graphed as a continuous series. Although the Cheyenne River graph is a composite of more than one type of stream, it serves to il-

lustrate the asymptotic nature of the rounding process.

Samples from the three creeks show the same general characteristics in their roundness and sphericity curves. In Rapid and Battle Creek curves the roundness increases rapidly at first and then more slowly. The roundness curve of Bear Butte Creek samples lacks the initial sharp increase observed in the other

TABLE 16

STATISTICAL PARAMETERS OF SHAPE AND ROUNDNESS DATA, MINNEKAHTA LIMESTONE, GRACE COOLIDGE CREEK AND BATTLE CREEK

| Sample No. | 16–32 Mm. Mean Roundness ($P$) | 16–32 Mm. Mean Sphericity ($\psi$) |
|---|---|---|
| C-A | 0.20 | 0.58 |
| C-B | 0.22 | 0.56 |
| C-C | 0.37 | 0.63 |
| C-D | 0.44 | 0.61 |
| B-1 | 0.52 | 0.60 |
| B-2 | 0.49 | 0.62 |
| B-3 | 0.53 | 0.61 |
| B-4 | 0.55 | 0.62 |
| B-5 | 0.55 | 0.61 |
| B-6 | 0.59 | 0.61 |

curves. As explained in the lithology section, the Minnekahta limestone pebbles of sample 1 in Bear Butte Creek are a mixture of pebbles from two sources, one of which is rather distant, so that rather high average roundness values are observed.

The limestone pebbles of Rapid Creek reach a higher roundness value for the distance traveled from their point of origin than do those of Battle Creek. At 22.5 miles the pebbles of Rapid Creek have a mean roundness of 0.65, as compared to 0.60 at the same distance in Battle Creek. The more rapid initial increase of roundness in Rapid Creek samples accounts for this. It has been shown in the lithology section that the limestone of Rapid Creek suffered more rapid attrition by abrasion than did the limestone of Battle Creek. This observation is further supported by the more

TABLE 17

STATISTICAL PARAMETERS OF SHAPE AND ROUNDNESS DATA, MINNEKAHTA LIMESTONE

| Sample No. | 16–32 Mm. Mean Roundness ($P$) | 16–32 Mm. Mean Sphericity ($\psi$) | 32–64 Mm. Mean Roundness ($P$) | 32–64 Mm. Mean Sphericity ($\psi$) |
|---|---|---|---|---|
| | Bear Butte Creek | | | |
| BB-1 | 0.51 | 0.62 | 0.50 | 0.60 |
| BB-2 | 0.52 | 0.59 | 0.53 | 0.65 |
| BB-3 | 0.55 | 0.66 | 0.55 | 0.61 |
| BB-4 | 0.56 | 0.63 | 0.57 | 0.64 |
| BB-5 | 0.64 | 0.63 | 0.64 | 0.65 |
| BB-6 | 0.64 | 0.62 | 0.63 | 0.64 |
| | Belle Fourche River | | | |
| BF-A | 0.71 | 0.63 | ......... | ......... |
| BF-B | 0.73 | 0.63 | ......... | ......... |
| | Cheyenne River | | | |
| Ch-A | 0.68 | 0.65 | ......... | ......... |
| Ch-B | 0.70 | 0.67 | ......... | ......... |
| Ch-C | 0.70 | 0.60 | ......... | ......... |
| Ch-D | 0.73 | 0.62 | ......... | ......... |
| Ch-E | 0.72 | 0.64 | ......... | ......... |

rapid rounding of Rapid Creek sediments, because rounding progresses mainly by abrasion.

Roundness is also a function of size. In the early stages the larger pebbles (32–64 mm.) are rounded at a more rapid rate than are the smaller ones (16–32 mm.). Scarcity of pebbles of large size beyond sample 6 made further comparisons impossible, but the data suggest

that the rates of rounding converge with distance. No definite conclusions can be drawn from a similar comparison of size to roundness in Bear Butte Creek.

The shape or sphericity of the Minnekahta limestone pebbles show similar characteristics in samples from all the creeks (figs. 18–20). The sphericity values fluctuate considerably in the first few miles and then remain fairly constant. The low values of sphericity encountered near the source are attributed to breakage rather than to selective transport, which would operate to increase the sphericity downstream.

One of the most important questions in a study of rounding of sedimentary particles is whether a particle, transported sufficiently far can reach a perfect roundness of 1.0 or whether some roundness value less than 1.0 is reached as a limit, depending on the composition, initial size and shape, and rigor of transport. The existing data on roundness suggest that the latter viewpoint is the correct one. Tumbling-barrel studies by Wentworth (1919, pp. 507–522) and Krumbein (1941*b*, pp. 482–520) indicate that rounding proceeds rapidly at first and then ever more slowly approaches an asymptotic value of roundness which is not exceeded. Field studies by these men support their laboratory conclusions. The present field study suggests that, in the case of limestone pebbles 16–32 mm., a limit of rounding of about 0.73–0.74 is

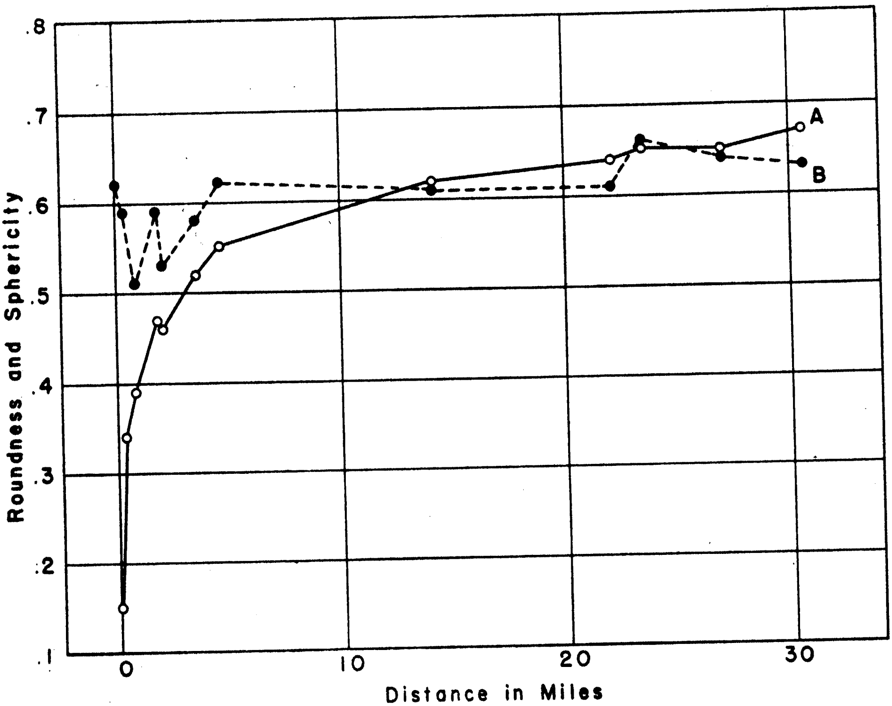



FIG. 18.—Relation of roundness and sphericity to distance of transport (Rapid Creek). Minnekahta limestone, 16–32 mm. *A*, roundness; *B*, sphericity.

approached after about 200 miles of transport. The roundness data of Wadell (1935, pp. 276–277) for various sizes of the St. Peter sandstone reveal surprisingly low roundness values for a sandstone generally considered to be very well rounded. Sand on the 0.5-mm. sieve possessed an average roundness of only 0.423.

Previous investigations have shown that, although sphericity and roundness are geometrically dissimilar properties, they are both functions of grain size and thus must be functions of each other. In general, sedimentary particles of high sphericity are rounder than those of low sphericity. Russell and Taylor (1937, pp. 225–267) established this correlation

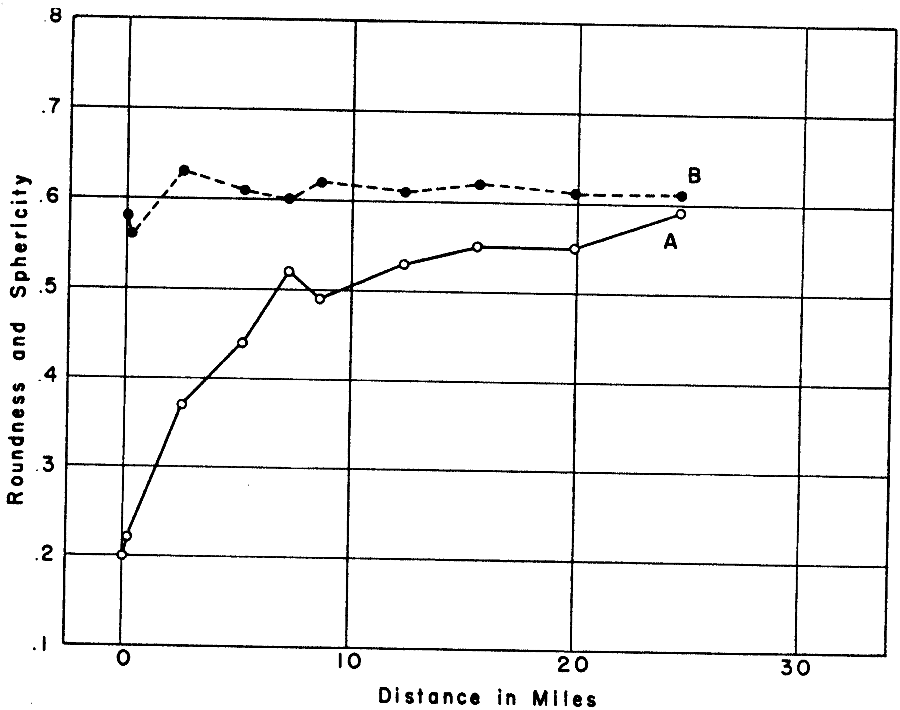



FIG. 19.—Relation of roundness and sphericity to distance of transport (Battle Creek). Minnekahta limestone, 16–32 mm. *A*, roundness; *B*, sphericity.

Reference to a limit of rounding which is not exceeded by a sedimentary particle, is, in reality, reference to the mean roundness of a sample of many particles. Individual pebbles or sand grains are commonly observed which have a roundness very close to 1.0, but the mean roundness of a sampled population of many particles is always less than 1.0.

for sands of the Mississippi River, and Pettijohn and Lundahl (1943, pp. 69–78) found similar relations to exist in Lake Erie beach sands. In both these studies roundness and sphericity were found to decrease with distance of transport, but roundness decreased at a greater rate than did sphericity. Plots of sphericity against roundness from the Lake Erie

data show a linear relationship. A 4.1 per cent increase in sphericity corresponded to a 21.2 per cent increase in roundness. Although these examples illustrate the relation between roundness and sphericity, they do not provide any explanation of how that relationship was established.

In figure 21 the average roundness of Black Hills sands has been plotted against average sphericity. Although the data scatter widely, a discernible trend is apparent toward higher roundness for increasing sphericity. For an increase of 4.7 per cent in sphericity, an increase of

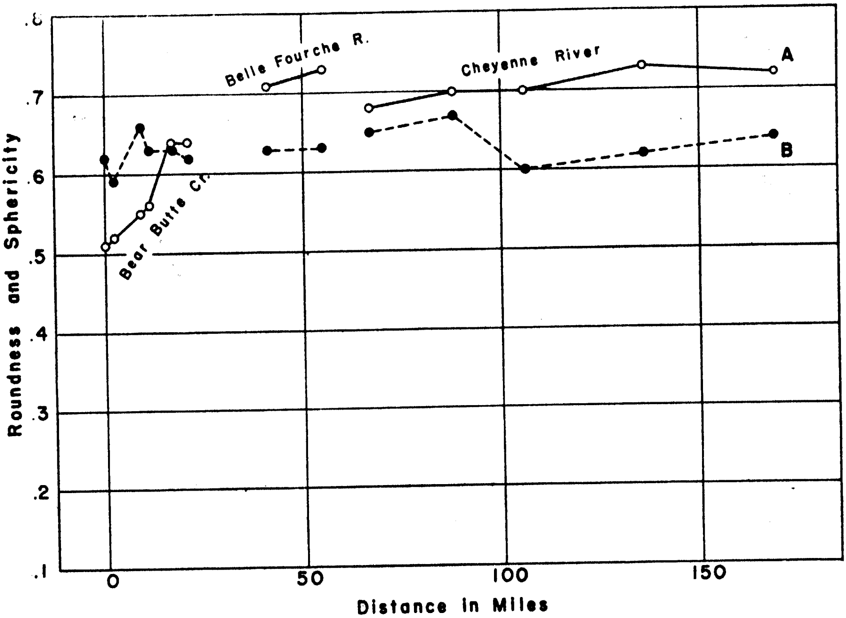



FIG. 20.—Relation of roundness and sphericity to distance of transport (Bear Butte Creek, Belle Fourche River, Cheyenne River). Minnekahta limestone, 16–32 mm. *A*, roundness; *B*, sphericity.

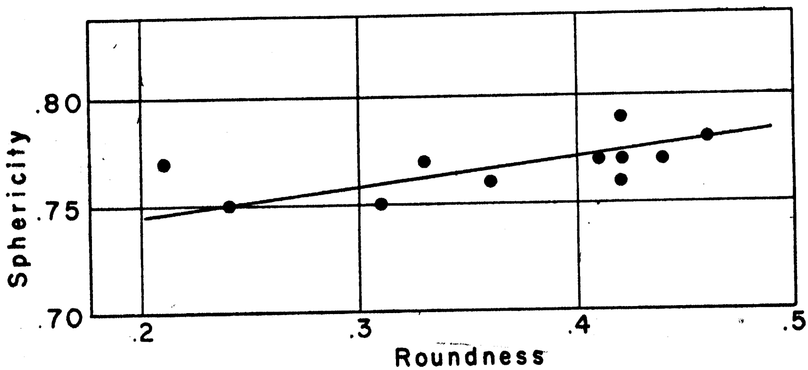



FIG. 21.—Relation of roundness and sphericity. Quartz sand, 1–1.414 mm. Arithmetic mean values of eleven samples of fifty grains each.

119.1 per cent is noted in the roundness. It is important to recognize that, although the relation of sphericity to roundness is probably basically the same in both the Lake Erie and the Black Hills sands, the actual percentage changes are not necessarily the same. Not only are the sizes studied somewhat different, but the Lake Erie samples represent a very heterogeneous mixture of sands from many sources with various histories. The decrease in their sphericity and accompanying decrease in roundness with distance of transport is apparently the result of selective transport, in which the less spherical grains in suspension outran the more spherical grains. On the other hand, the increase in sphericity and roundness of the Black Hills sands is apparently the result of abrasion. Figure 21 illustrates the much greater effect of transportation on the roundness of sand grains than on their sphericities.

Figure 22 illustrates the relation between roundness and sphericity for 16–32-mm. limestone pebbles. A 12.3 per cent increase in sphericity is accompanied by a 380 per cent increase in roundness. The ratios of sphericity increase to roundness increase are roughly the same for both the quartz sand grains and the limestone pebbles. This fact suggests that the rate of sphericity increase, as compared to the rate of roundness increase, is about the same for particles of widely different size and composition.

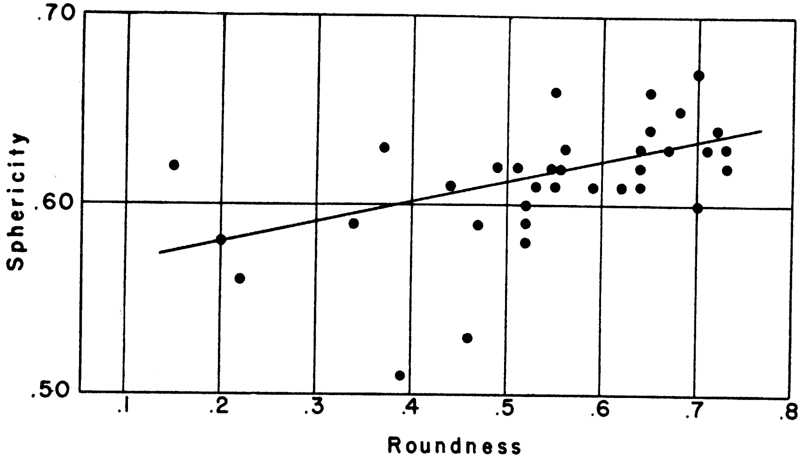



FIG. 22.—Relation of roundness and sphericity. Minnekahta limestone, 16–32 mm. Arithmetic mean values of thirty-five samples of fifty pebbles each.

## SAND ANALYSIS

### PROBLEMS OF SAMPLING

Sand samples from the present-day streams were collected to study the effects of transportation on small particles. The samples were limited to the 1–1.414-mm. size to eliminate contamination from outside sources, once the original source area was left behind. The source for this sand is the Harney Peak granitic area of the southern Black Hills. This area furnishes coarse arkosic sand to streams such as Battle Creek and eventually the Cheyenne River. Because of the coarseness of the sand, contamination of the samples from the finer-grained

Paleozoic and Mesozoic sandstones surrounding the Hills is negligible. This assumption is supported by Dake (1921, chap. ii) who studied the textures of a large number of sandstones. He found that in such sandstones the percentage of quartz grains in excess of 0.6 mm. was quite insignificant (0.1–0.3 per cent). Thus any textural changes in these sand samples are due to the effects of transportation.

### LABORATORY METHODS

Each sand sample was quartered with an Otto microsplit (Otto, 1933, pp. 30–39). The sand was examined with a binocular microscope, and the grains unmistakably quartz were separated from the feldspar grains plus doubtful grains. The latter fraction was crushed in a mortar, grain by grain, and examined in oil with a petrographic microscope. This process was necessary because some quartz grains are impossible to distinguish from feldspar megascopically and staining techniques proved unreliable. One hundred and fifty grains were examined in this way for each sample. The probable error is estimated at 9 per cent (Krumbein and Pettijohn, 1938, p. 472).

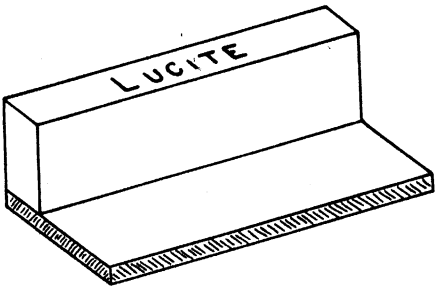



FIG. 23.—Mounting device for determining intercept sphericities of sand grains.

Shape and roundness determinations were confined to quartz grains (1–1.414 mm.) only. Fifty grains from each sample were mounted in rows on microscope slides with gum arabic and were then projected with a microprojector. The roundness was determined visually, as in the pebble analysis. The intercept method was applied to determine sphericity. A special mount was required for measurement of the short axis of the grain. This required a microscope slide and lucite block to which the slide was cemented, as pictured in figure 23. The grains are mounted in a row along one edge of the glass slide, so that their planes of maximum area are approximately parallel to the surface of the slide. The mount is then placed on a microscope stage, and the long and intermediate diameters are measured with a micrometer ocular. The slide is then turned on edge, so that it rests on the lucite block. The short diameter can then be measured directly with the micrometer ocular.

### ANALYSIS AND PRESENTATION OF DATA

The statistical parameters of shape, roundness, and composition are listed in table 18. In figures 24 and 25 these data are plotted against distance as the independent variable.

*Feldspar content.*—The feldspar content decreases rapidly downstream in Battle Creek, a loss of 51.2 per cent in 40 miles (fig. 24). In the Cheyenne River the loss of feldspar is 13.8 per cent in 150 miles. The sand contribution of a stream such as Battle Creek must be small in comparison to the volume of sand carried by the Cheyenne River, because the higher feldspar content of the latter is apparently little affected by tributary contributions. The high feldspar content of the Cheyenne River in comparison to that of Battle Creek is a reflection of the source of the feldspar. Farther upstream on the Cheyenne River, above the mouth of Battle Creek, the tributaries are pro-

gressively shorter. They also drain areas progressively richer in feldspar. These areas are rich in pegmatites containing as much as 60 per cent feldspar. Thus these tributaries start with sand having a high feldspar content and transport it shorter distances before entering the Cheyenne River. The combination of these two factors results in the Cheyenne River's having, at any point, a higher feldspar content than does a tributary stream entering it at that point.

R. D. Russell (1937, pp. 1307–1348) took issue with the accepted idea that certain minerals, such as feldspar, are rapidly eliminated during stream transport. In the 100-mesh grade (0.147–0.208 mm.) of Mississippi River sands, he found that the feldspar content decreased 20 per cent in 1,100 miles of transport. However, on the basis of a few whole samples (all grades), the feldspar content increased slightly downstream. He concluded that destruction of feldspar by abrasion and alteration during transport is offset by other factors, such as selective sorting on the basis of size, the breakage of larger grains thus adding to the feldspar content of the smaller sizes.

Although only one size was studied in the Black Hills sands, the data suggest that wear is much more effective in reducing feldspar in streams such as Battle Creek and the Cheyenne River than it is in the Mississippi River. The decline in feldspar in the Black Hills streams can be attributed only to a combination of abrasion and weathering during transport. Contamination is absent, as is any selective sorting on the basis of size. The results are comparable to those obtained by Mackie (1896, pp. 148–172), who found that the feldspar content of certain streams in Scotland decreased as much as 50 per cent in 30–40 miles. It is concluded that in streams similar to Battle Creek and the Cheyenne River, the feldspar content of coarse sand is rapidly reduced by the effects of transportation.

*Shape and roundness*.—The question of how rapidly and under what conditions sand grains become rounded has long been perplexing. In a recent article Twenhofel (1945, pp. 59–71) summarized the arguments. He concluded, on the basis of the evidence at hand, that rounding of sand grains is largely, if not entirely, done in traction transport. Furthermore, he concluded that quartz grains of sand dimensions are very little, if at all, rounded in streams, especially in high-velocity streams, where the general tendency is in the direction of increasing angularity. Contrary to this view, Krynine (1940, p. 81) reported that in the initial stage of transport the passage from angular to subangular sand grains is ac-

TABLE 18

STATISTICAL PARAMETERS OF SHAPE, ROUNDNESS, AND COMPOSITION, SIZE CLASS 1–1.414 MM., BATTLE CREEK AND CHEYENNE RIVER

| Sample No. | Mean Roundness ($P$) | Mean Sphericity ($\psi$) | Per Cent Feldspar |
|---|---|---|---|
| | Battle Creek | | |
| SS-1 | 0.21 | 0.77 | 35 |
| SS-2 | 0.24 | 0.75 | 29 |
| SS-3 | 0.31 | 0.75 | 23 |
| SS-4 | 0.33 | 0.77 | 20 |
| SS-5 | 0.36 | 0.76 | 17 |
| | Cheyenne River | | |
| SS-6 | 0.42 | 0.76 | 29 |
| SS-7 | 0.41 | 0.77 | 30 |
| SS-8 | 0.42 | 0.77 | 27 |
| SS-9 | 0.42 | 0.79 | 26 |
| SS-10 | 0.46 | 0.78 | 24 |
| SS-11 | 0.44 | 0.77 | 25 |

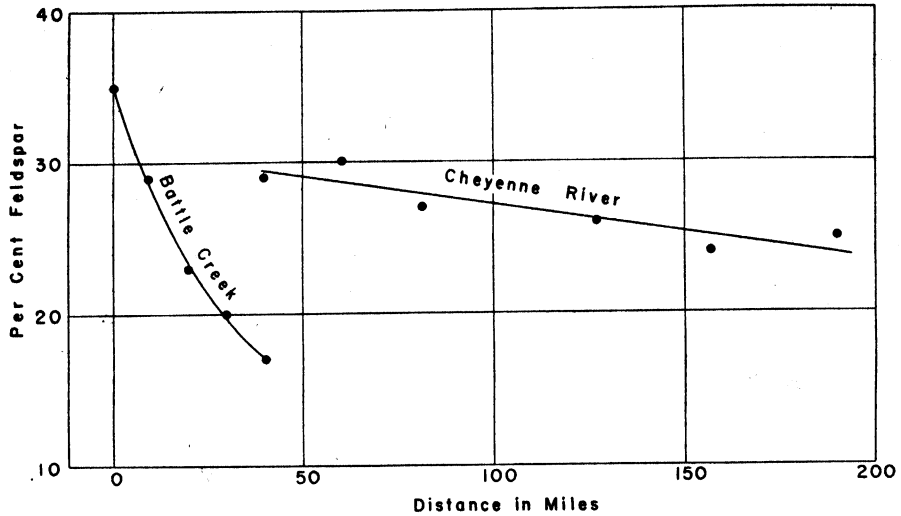



FIG. 24.—Relation of percentage of feldspar to distance of transport (Battle Creek and Cheyenne River) Size class, 1–1.414 mm.

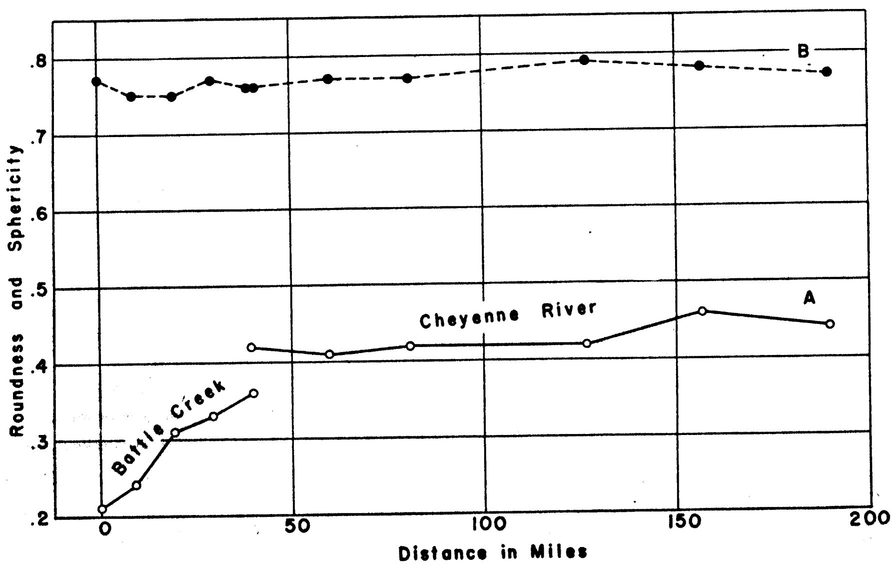



FIG. 25.—Relation of roundness and sphericity to distance of transport (Battle Creek and Cheyenne River). Quartz sand, 1–1.414 mm. *A*, roundness; *B*, sphericity.

complished in a very short period. The Battle Creek roundness curve (fig. 25) shows this to be true in quartz sand (1–1.414 mm.). The grains in the Cheyenne River show a continued but much slower increase of roundness with distance. The higher roundness values of the Cheyenne River samples reflect the greater distance of transport that the grains have experienced in that river before reaching the mouth of Battle Creek.

The change of sphericity with distance is shown in figure 25. In Battle Creek the sphericity values are too irregular for one to discern any trend. In the Cheyenne River a slight increase in sphericity with distance is noted.

These results are different from those found by Russell and Taylor (1937, pp. 225–267). They found that in 1,100 miles of transport in the Mississippi River the arithmetic mean roundness of the sizes (0.074–0.417 mm.) decreased from 0.234 to 0.179. In the same distance the sphericity decreased from 0.825 to 0.809. The decrease in roundness was attributed to breakage primarily, although selective sorting on the basis of shape would account for the same result. It is a curious paradox that breakage of sand grains should be so prevalent in the Mississippi River and apparently absent in a high-velocity stream like Battle Creek. The rounding of quartz sand grains in Battle Creek can be accounted for only by abrasion. Because only quartz grains were studied, change in mineral composition is not a factor, nor is contamination from other sand sources, since there are none in that size range. Selective sorting on the basis of shape is not a factor, inasmuch as no discernible sphericity trends are present.

For the above reasons, it is concluded that the increase in grain roundness observed in the Cheyenne River is probably due to abrasion also. It is possible that here the increase may be due to a selective sorting on the basis of shape, as the sphericity rise is about equal to the roundness increase. However, the rounding of grains in Battle Creek was seen to be completely independent of sphericity change, so it is not necessary to rely on sphericity change to explain rounding in the Cheyenne River. Abrasion alone is adequate.

### EQUATION OF ROUNDING

Sedimentary particles have been observed to become rounder when transported by streams. This has been proved for particles of pebble size by many investigators. Data from Battle Creek show that streams also round sand grains as small as 1–1.414 mm. The first quantitative measure of the rate of rounding for pebbles was made by Wentworth (1919, pp. 507–522), who found that rounding proceeds rapidly at first and then less rapidly with distance. Krumbein (1940, pp. 639–676) expressed this relationship with the following differential equation:

$$\frac{dP}{dx} = k\,(P_0 - P)\,, \qquad (a)$$

which states that the rate of change of roundness ($P$) with distance ($x$) is equal to the difference between the roundness at any point and some limiting value of roundness ($P_0$), multiplied by a proportionality constant $k$. Solution of this differential equation gives:

$$P = P_0(1 - e^{-kx})\,, \qquad (b)$$

which shows the exponential nature of the relationship. This equation adequately fits the roundness data from San Gabriel Canyon.

Later tumbling-barrel experiments by Krumbein (1941*b*, pp. 482–520) showed

that the function was more complicated. Roundness plotted against distance on semilog paper did not give a straight line, as required for an exponential function of this type. Sphericity plotted as an essentially straight line, but roundness showed an initial steep curve and a curve which paralleled the sphericity curve after about 5 miles.

Krumbein developed the following differential equation to fit the roundness curve:

$$\frac{d(P-P_0)}{ds} = b(\psi - \psi_0) - a_2(P-P_0) . \quad (c)$$

Solution of this equation gives:

$$P_0 - P = (P_0 - P_i)\, e^{-a_2 s} + \frac{b(\psi_0 - \psi_i)}{a_2 - a_3} \times [e^{-a_3 s} - e^{-a_2 s}] , \quad (d)$$

where $P_i$ = initial roundness, $\psi_0$ = limiting sphericity, $\psi_i$ = initial sphericity, $a_2$ and $a_3$ = coefficients of rounding and sphericity, respectively, and $b$ = shape coefficient in rounding.

Krumbein contended that the initial steep part of the rounding curve is controlled by the first term in equation (*d*). The long-term aspects of rounding, however, are controlled by the rate of sphericity increase, as shown by the second term in equation (*d*). Thus Krumbein found that, for certain abrasive conditions in a tumbling barrel, the long-term aspects of rounding are governed by the initial sphericity of the particle and the rate of sphericity increase.

Similar, but more extensive, experiments were later made by Sarmiento (1945, pp. 1–59). Limestone pebbles of three sizes (16, 32, and 64 mm.) were abraded in a tumbling barrel, with and without sand, each size alone and also in mixed sizes. In general, roundness increased, as found by Krumbein. However, sphericity change was very erratic. When each size was abraded separately, the sphericity behaved irregularly, either showing no change or a slight decrease. When the three sizes were abraded together, the sphericities of the 32- and 64-mm. sizes decreased slightly and that of the 16-mm. size markedly (0.69–0.48 in 20 miles). This sphericity decrease was ascribed to breakage. Sarmiento concluded that roundness changes are independent of sphericity changes.

The Black Hills field data tend to support Sarmiento's laboratory conclusion. Graphs of pebble sphericity plotted against distance (figs. 18–20) show no obvious systematic increase in sphericity. The initial sphericity irregularities have been discussed previously.

It is apparent that neither of the assumptions expressed in Krumbein's differential equations (*a*) and (*c*) is quite correct. In the following treatment it has been assumed that the rate of change of roundness with distance is not only proportional to the difference between the roundness at any point and a limiting roundness but also to some power of the distance traveled.

Expressed mathematically:

$$\frac{dP}{dS} = KS^a(P_L - P) , \quad (1)$$

where $P$ = roundness, $S$ = distance, $K$ = constant of proportionality, $a$ = a negative constant, and $P_L$ = limiting value of roundness.

To avoid confusion in symbolism, $P_L$ is used for the limiting value of roundness instead of $P_0$, which in most exponential notations indicates an initial value.

To account for not starting where $P = 0$, let $S = x + x_0$, where $S = 0$ when $P = 0$. To simplify the mathematics and form of the solution, let

$K = nr$ and $\sigma = n - 1$, where $n$ and $r$ are constants. Then,

$$\frac{dP}{dx} = n\,r\,(x + x_0)^{n-1}(P_L - P)\,. \quad (2)$$

Solution of this differential equation follows:

$$\frac{dP}{P_L - P} = n\,r\,(x + x_0)^{n-1}dx\,.$$

Then, integrating, we obtain:

$$-\log_e(P_L - P) = r\,(x + x_0)^n + C\,.$$

The parameters of this equation are $n$ and $r$, which can be obtained graphically by converting equation (3) to the log base 10, as follows:

$$-2.3\,\log_{10}\frac{P_L - P}{P_L} = r\,(x + x_0)^n\,.$$

Take the log of both sides, then

$$\log_{10}\left(-2.3\,\log_{10}\frac{P_L - P}{P_L}\right) = \log_{10} r + n\,\log_{10}(x + x_0)\,. \quad (5)$$

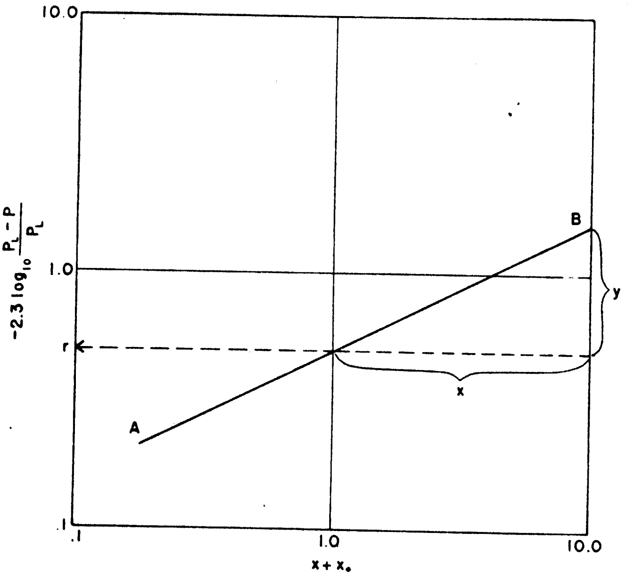



FIG. 26.—Graphical solution for the constants $n$ and $r$ of the rounding equation

To evaluate $C$, at $x + x_0 = 0$, $P = 0$. Then

$$-\log_e P_L = C\,.$$

Substitution of the value of $C$ gives

$$\log_e\frac{P_L - P}{P_L} = -r\,(x + x_0)^n\,, \quad (3)$$

$$\frac{P_L - P}{P_L} = e^{-r(x+x_0)^n}\,.$$

The solution of the differential equation thus is:

$$P = P_L\,[1 - e^{-r(x+x_0)^n}]\,. \quad (4)$$

This equation is of the form $y = r + nx$, which may be plotted on logarithmic paper (fig. 26). The slope of the line $AB$ is $n$ and equals $y/x$ by direct measurement. To find $r$, let $\log_{10}(x + x_0) = 0$. Then $r = -2.3\log_{10}(P_L - P)/P_L$ at the intersection of line $AB$ and $x + x_0 = 1.0$.

The rounding of pebbles observed in two creeks and in one tumbling-barrel experiment has been treated on the basis of the above analysis. Calculation sheets are shown in tables 19–21.

Certain assumptions must be made with regard to the values of $x_o$ and $P_L$ in each calculation. The quantity $x_o$ represents the theoretical distance that a pebble would have traveled to increase its roundness from zero to the value found in the outcrop. However, no transport is actually implied here, because that initial rounding may be due to weathering. For any given case, the choice of $x_o$ depends on the rate of initial rounding, which in the cases cited was highest for Rapid Creek and lowest for Battle Creek, with Sarmiento's experiment falling between the two. The choice of $P_L$ considerably affects the other extreme of the curve. Choices of $P_L$ from 0.02 to 0.05 higher than the last observed roundness value were found to be most satisfactory.

The data have been plotted on logarithmic paper in figure 27. This graph determines the parameters $r$, $n$, $K$, and $a$ in table 22.

The constant $r$ is the coefficient of rounding. However, it is not identical

TABLE 19

COMPUTATION CHART (FOR DETERMINATION OF PARAMETERS $r$ AND $n$), RAPID CREEK (MINNEKAHTA LIMESTONE PEBBLES, 16–32 MM.)

($x_o$ = 0.01 Miles; $P_L$ = 0.72)

| Sample No. | $x$ (Miles) | $x+x_o$ (Miles) | $P$ | $P_L-P$ | $(P_L-P)/P_L$ | $\log_{10}(P_L-P)/P_L$ | $-2.3\log_{10}(P_L-P)/P_L$ |
|---|---|---|---|---|---|---|---|
| R-A | 0 | 0.01 | 0.15 | 0.57 | 0.79 | −0.1024 | 0.236 |
| R-B | 0.3 | 0.31 | 0.34 | 0.38 | 0.53 | −0.2757 | 0.635 |
| R-1 | 0.8 | 0.81 | 0.39 | 0.33 | 0.46 | −0.3372 | 0.777 |
| R-C | 1.8 | 1.81 | 0.47 | 0.25 | 0.35 | −0.4559 | 1.050 |
| R-D | 2.0 | 2.01 | 0.46 | 0.26 | 0.36 | −0.4437 | 1.020 |
| R-2 | 3.5 | 3.51 | 0.52 | 0.20 | 0.28 | −0.5528 | 1.277 |
| R-3 | 4.6 | 4.61 | 0.55 | 0.17 | 0.24 | −0.6198 | 1.425 |
| R-6 | 14.1 | 14.11 | 0.62 | 0.10 | 0.14 | −0.8537 | 1.880 |
| R-7 | 22.1 | 22.11 | 0.64 | 0.08 | 0.11 | −0.9586 | 2.200 |
| R-E | 23.5 | 23.51 | 0.65 | 0.07 | 0.10 | −1.0000 | 2.300 |
| R-8 | 27.0 | 27.01 | 0.65 | 0.07 | 0.10 | −1.0000 | 2.300 |
| R-10 | 30.6 | 30.61 | 0.67 | 0.05 | 0.07 | −1.1549 | 2.540 |

TABLE 20

COMPUTATION CHART (FOR DETERMINATION OF PARAMETERS $r$ AND $n$), BATTLE CREEK (MINNEKAHTA LIMESTONE PEBBLES 16–32 MM.)

($x_o$ = 0.5 Miles; $P_L$ = 0.65)

| Sample No. | $x$ (Miles) | $x+x_o$ (Miles) | $P$ | $P_L-P$ | $(P_L-P)/P_L$ | $\log_{10}(P_L-P)/P_L$ | $-2.3\log_{10}(P_L-P)/P_L$ |
|---|---|---|---|---|---|---|---|
| C-A | 0 | 0.5 | 0.20 | 0.45 | 0.69 | −0.1612 | 0.370 |
| C-B | 0.2 | 0.7 | 0.22 | 0.43 | 0.66 | −0.1805 | 0.415 |
| C-C | 2.1 | 2.6 | 0.37 | 0.28 | 0.43 | −0.3665 | 0.843 |
| C-D | 4.5 | 5.0 | 0.44 | 0.21 | 0.32 | −0.4949 | 1.14 |
| B-1 | 6.5 | 7.0 | 0.52 | 0.13 | 0.20 | −0.6980 | 1.60 |
| B-2 | 7.8 | 8.3 | 0.49 | 0.16 | 0.25 | −0.6021 | 1.39 |
| B-3 | 11.4 | 11.9 | 0.53 | 0.12 | 0.18 | −0.7447 | 1.72 |
| B-4 | 14.5 | 15.0 | 0.55 | 0.10 | 0.15 | −0.8239 | 1.89 |
| B-5 | 18.5 | 19.0 | 0.55 | 0.10 | 0.15 | −0.8239 | 1.89 |
| B-6 | 22.6 | 23.1 | 0.59 | 0.06 | 0.09 | −1.0458 | 2.41 |

with either $k$ or $a_2$ of Krumbein's equations ($a$) and ($b$), since it is equal to $K/n$. Only if $n = 1$ or $a = 0$ does $r$ equal Krumbein's $k$.

The use of $n$ permits the roundness data to be plotted as a straight line on semilog paper, where the ordinate is $(P_L - P)/P_L$ and the abcissa is some power ($n$) of the distance $(x + x_0)$. In Rapid Creek $n$ was found to be about $\frac{1}{3}$; in the other cases, $\frac{1}{2}$.

Thus the history of the rounding of

TABLE 21

COMPUTATION CHART (FOR DETERMINATION OF PARAMETERS $r$ AND $n$) SARMIENTO'S TUMBLING-BARREL, SECOND EXPERIMENT (LIMESTONE PEBBLES, 32 MM.)

($x_0$ = 0.2 Miles; $P_L$ = 0.74)

| Sample No. | $x$ (Miles) | $x-x_0$ (Miles) | $P$ | $P_L-P$ | $(P_L-P)/P_L$ | $\log_{10}(P_L-P)/P_L$ | $-2.3\log_{10}(P_L-P)/P_L$ |
|---|---|---|---|---|---|---|---|
| 1 | 0 | 0.2 | 0.21 | 0.53 | 0.72 | −0.1427 | 0.328 |
| 2 | 0.17 | 0.39 | 0.22 | 0.52 | 0.70 | −0.1549 | 0.357 |
| 3 | 0.33 | 0.53 | 0.27 | 0.47 | 0.64 | −0.2007 | 0.446 |
| 4 | 0.50 | 0.70 | 0.31 | 0.43 | 0.58 | −0.2366 | 0.544 |
| 5 | 1.0 | 1.2 | 0.36 | 0.38 | 0.51 | −0.2925 | 0.673 |
| 6 | 2.0 | 2.2 | 0.46 | 0.28 | 0.38 | −0.4203 | 0.969 |
| 7 | 3.0 | 3.2 | 0.49 | 0.25 | 0.34 | −0.4686 | 1.08 |
| 8 | 4.0 | 4.2 | 0.50 | 0.24 | 0.32 | −0.4949 | 1.14 |
| 9 | 5.0 | 5.2 | 0.56 | 0.18 | 0.24 | −0.6198 | 1.42 |
| 10 | 7.0 | 7.2 | 0.59 | 0.15 | 0.20 | −0.6980 | 1.61 |
| 11 | 9.0 | 9.2 | 0.60 | 0.14 | 0.19 | −0.7213 | 1.66 |
| 12 | 12.0 | 12.2 | 0.64 | 0.10 | 0.135 | −0.8697 | 2.00 |
| 13 | 16.0 | 16.2 | 0.67 | 0.07 | 0.095 | −1.0223 | 2.35 |
| 14 | 20.0 | 20.2 | 0.70 | 0.04 | 0.054 | −1.2677 | 2.92 |
| 15 | 30.0 | 30.2 | 0.71 | 0.03 | 0.040 | −1.3970 | 3.21 |
| 16 | 40.0 | 40.2 | 0.72 | 0.02 | 0.027 | −1.5870 | 3.65 |

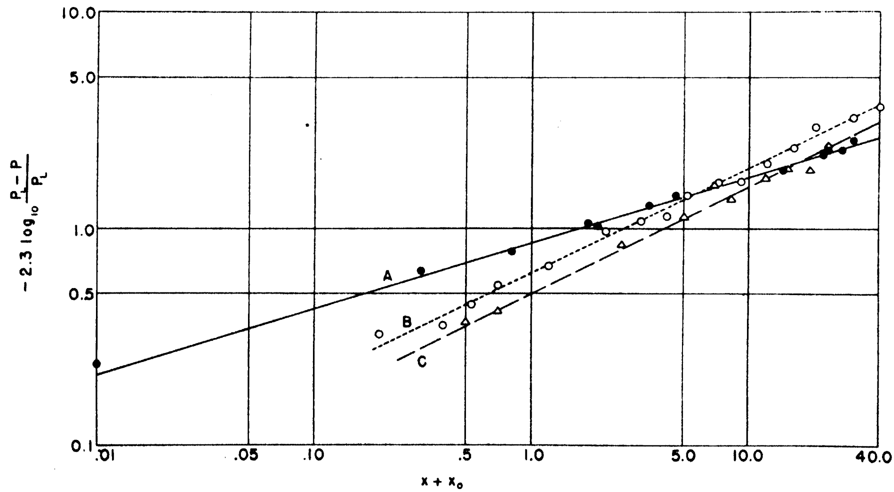



FIG. 27.—Logarithmic plot of rounding equation data. $A$, Rapid Creek; $B$, Sarmiento's second experiment; $C$, Battle Creek.

particles in a stream can be expressed by two parameters and by an equation, which, in the apparent absence of any sphericity effect, more correctly expresses the factors involved.

## GEOLOGICAL INFERENCES

### ABRASION AND SELECTIVE TRANSPORT

As previously stated, size reduction in a stream may be effected by two processes: (1) selective transport and (2) abrasion and breakage. Hitherto it has been impossible to evaluate these two processes quantitatively.

TABLE 22

NUMERICAL CONSTANTS OF THE ROUNDING EQUATION

| | $r$ | $n$ | $K=nr$ | $a=n-1$ |
|---|---|---|---|---|
| Rapid Creek (16–32 mm.) | 0.86 | 0.31 | 0.266 | −0.69 |
| Sarmiento 2d experiment, 32 mm. | 0.62 | 0.50 | 0.310 | −0.50 |
| Battle Creek (16–32 mm.) | 0.50 | 0.50 | 0.250 | −0.50 |

The competence of a stream to carry a certain size of material is a function of stream discharge and gradient. As the gradient of a stream decreases away from its headwaters, the mean size of sediment carried is decreased. The correlation between mean size and stream gradient has been shown in figure 10. Thus the size decrease shown in the Black Hills terrace deposits must be mainly a function of selective transport. However, part of this size decrease is certainly due to abrasion and breakage. Pebbles become rounder in stream transport, which is proof that abrasion is taking place.

The lithology data (table 13) furnish a means of computing the proportion of size decrease due to abrasion. In sample R-1 (Rapid Creek) the rock types may be divided into two groups: the hard rocks (chert, quartz, quartzite) and the soft rocks (sandstone, limestone, pre-Cambrian metamorphics). The hard rocks constitute about 40 per cent of sample R-1 and the soft rocks 60 per cent. In a distance of 30 miles (sample R-10) the hard rocks constitute about 90 per cent of the sample and the soft rocks 10 per cent (recomputed to 100 per cent). Because it is unlikely that the decrease in soft rocks is due to any selective action, it must be accounted for by abrasion. The number percentages may be regarded as volume percentages. Thus in 30 miles of transport the volume of soft rocks has decreased from 60 to 10 per cent, a reduction of 50 per cent in a unit volume made up of both soft and hard rocks. A reduction of 50 per cent in a unit volume of a sphere results in a 20 per cent decrease in diameter. Therefore, on the basis of abrasion alone, one would expect a size reduction of 20 per cent in the mean diameter of the gravel in 30 miles. However, the size data from Rapid Creek reveal a decrease in that distance from about 24-mm. diameter to 5-mm. diameter, a reduction of about 80 per cent. The conclusion is that selective transport accounts for 75 per cent of the size reduction observed in Rapid Creek and abrasion for the remaining 25 per cent.

Similar calculations for Battle Creek reveal that 84 per cent of the size reduction is due to selective transport and 16 per cent to abrasion. Evidence furnished by the rates of rounding of limestone pebbles gives further proof that abrasion is less effective in Battle Creek than in Rapid Creek. Figures 18 and 19 illustrate this fact.

It has been assumed that the loss of the 16–32-mm. soft rocks by abrasion is an average loss for the entire range of

sizes. This assumption is believed to be justified by the fact that the average size of all channel samples is 12 mm., only slightly less than the size studied.

### INDICES OF MATURITY

The main effort in this work has been directed toward a solution of some of the problems involved in the transportation of sedimentary material in streams. An attempt has been made not only to measure quantitatively the action of stream transport but also to define the fundamental principles which govern the measured effects.

Quantitative measurements, such as change of roundness or feldspar content of a sand with distance of transport, enable one to derive the relationships involved in such changes. Knowledge of these relationships provides a basis for extrapolating the quantitative data beyond the scope of observation and experimentation.

The importance of this possibility of extrapolation appears in consideration of what may be called "indices of maturity." Three fundamental "indices" are roundness, sphericity, and lithology. On the basis of these "indices" a clastic deposit may be classified as to its degree of maturity. At the extreme youthful end of the scale a clastic sediment is characterized by extreme angularity and relatively low sphericity of the component particles and a high percentage of chemically and mechanically unstable constituents. At the other extreme of old age a clastic sediment is composed of particles having a high roundness and high sphericity and lacking any appreciable percentage of chemically and mechanically unstable constituents.

The concept of maturity as here expressed is concerned not only with the stage of maturity reached by any sediment but also with the effectiveness of different geologic agents in producing that stage of maturity. The Black Hills data provide a means of calculating the effects of stream transport on some of these "indices of maturity."

*Sand roundness.*—Sand from the vicinity of Harney Peak represents material close to the youthful end of the maturity scale. It is very angular and has a high percentage of feldspar. The other end of the scale is represented by the St. Peter sandstone (Ordovician), a very pure, well-rounded quartz sand (averaging above 98.0 per cent silica). Wadell (1935, pp. 276–277) computed the roundness of several size grades of samples from this sandstone. His data have been plotted on logarithmic paper in figure 28. These data plot as a straight line and illustrate the relation between roundness and grain size. The size-roundness relation has been extrapolated to larger sizes by the writer by means of a dashed line. The extrapolation indicates that sand grains 1–1.414 mm., like those studied in the Black Hills sands, would possess a roundness of about 0.5. Thus it is indicated that the roundness of sand grains in a very mature sand is relatively low, further evidence that a limit of roundness is reached.

If a roundness value of 0.5 is assumed to be the limit of rounding for quartz grains 1–1.414 mm. in size, one can calculate the distance such a particle would have to be transported in a stream to reach this limiting value. The equation of rounding (4) developed in a preceding section was applied to the roundness-distance data of quartz sand (table 14). The constant $n$ in this equation was found to equal 0.64 and $r$ to equal 0.10. It was assumed that the limit of rounding is 0.5 and that the roundness at $x$ miles is 0.499. Solution of this equation for $x$ re-

veals that, to round a 1–1.414-mm. quartz sand grain to a roundness of 0.499, the grain would have to be transported a distance of about 600 miles. Half this rounding is accomplished in the first 20 miles. It is apparent that for very coarse quartz sand a mature index of rounding is reached in a surprisingly short distance of stream transport. On the basis of data available at present it is not possible to calculate the distance of transport required to round to their limit sands of smaller size. Wadell's data (fig. 28) reveal

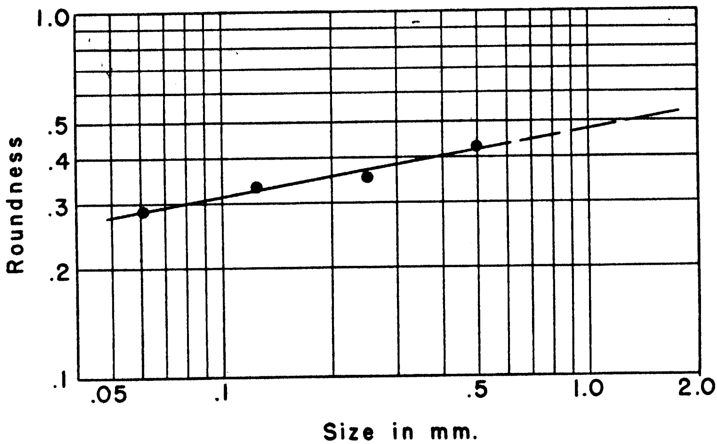



FIG. 28.—Relation of log arithmetic mean roundness and log geometric mean diameter. St. Peter sandstone (Wadell).

that the smaller sizes have correspondingly lower maturity indices of roundness. Since the process of rounding is a function of size, it is probable that sand grains less than 1 mm. would require far greater distances of transport to reach their limits of rounding. Unfortunately, Thiel's experiments on abrasion of sand grains furnished no quantitative roundness data. Experiments of that type, but which include quantitative roundness measures, should provide data for approximating the distances required to round quartz grains of small size to their limits of rounding.

*Sand sphericity.*—Studies do not establish sphericity as a reliable index of maturity. For it to be useful in this respect the sphericity of a youthful sand should be noticeably different than that of a mature sand. Some evidence on this question is found by comparing sphericities of quartz grains from the very youthful sands of the Black Hills with those of the mature St. Peter sandstone. The intercept sphericities of the former average about 0.76, while Wadell's data from the St. Peter sandstone for sphericities of grains of comparable size average about 0.86. Wadell's sphericities, however, are projection sphericities, which err in being too high as compared to intercept sphericities. Comparison of intercept with projection sphericities of the same samples from the data of Pettijohn and Lundahl (1943, pp. 69–78) indicates that the latter average about 0.07 too high. Thus it is indicated that the sphericity index of a youthful sand is too close to that of a mature sand to be of any practical value. Corroborative evidence on this conclusion is supplied by Ingerson and Ramisch (1942, pp. 595–606). They found that

elongation of sand grains parallel to the *c*-axis (i.e., St. Peter sandstone) was a characteristic inherited from the parent igneous and metamorphic source rocks and not the result of abrasion and fracturing of the grains during transport.

On the other hand, Thiel's work (1940, pp. 103–124) offers evidence that sphericity change with abrasion is pronounced enough to provide a useful maturity index. He found that in 5,000 miles of transport the sphericity of quartz sand 1–2 mm. increased from 0.72 to 0.79. Three factors are important in evaluating the significance of this change. First, projection sphericities were measured and thus cannot be compared directly to true sphericity changes. Second, the sphericities of the artificially crushed quartz are considerably lower than those of natural quartz particles resulting from the breakup of igneous and metamorphic rocks. Third, the sphericity increased six times as fast in the last 3,000 miles as it did in the first 2,000 miles. This result is hard to explain on the basis of known facts of sphericity change, which would lead one to expect the greatest sphericity increase in the early stages of abrasion. Because Thiel's data contain factors unlike those of a natural environment and other evidence on sphericity change is contradictory to his conclusions, it is concluded that sphericity does not furnish a good index of maturity.

*Unstable constituents of sand.*—In this category, as an index of maturity, may be listed the percentages of feldspar and those heavy minerals which are mechanically and chemically unstable. Because the sand samples from the Black Hills were of a size too large to contain heavy minerals, the discussion will be limited to the feldspar percentage.

The graph of feldspar decrease with distance (fig. 24) suggests that the rate of feldspar decrease is a function of the stream gradient. The rapid feldspar decrease in Battle Creek is associated with the steep gradient of a mountain stream. A much slower decrease is associated with a gentler gradient of the Cheyenne River. Plots of these data on semilog paper reveal essentially straight lines. Thus, as a first approximation, it can be assumed that the rate of feldspar decrease follows an exponential law. Since data on feldspar decrease (1–1.414 mm.) are not available downstream in the Missouri and Mississippi rivers, it is possible to arrive at only a rough approximation of their effects on material of this size. If it were assumed that the exponential relation of feldspar decrease for the Cheyenne River held for the Missouri and Mississippi rivers and that no more feldspar was added, 1–1.414-mm. sand would have its feldspar content reduced to 14 per cent by the time it reached the Gulf of Mexico. Because the gradients of the Missouri and Mississippi rivers are much less than that of the Cheyenne River, it is probable that the feldspar decrease is not nearly so rapid as the above approximation indicates. R. D. Russell (1937, pp. 1307–1348) found that for somewhat smaller size grades in the Mississippi River the feldspar content varied irregularly with distance of transport. Some grades showed a feldspar decrease and others an increase. Thus it is apparent that, in spite of the rapid feldspar decrease of a sand near its source, one cycle of stream transport is not sufficient to produce an old-age feldspar index.

*Gravel deposits.*—Indices of maturity may also be applied to gravel deposits. Since abrasion affects large particles to a greater extent than it does small ones, it is to be expected that mature indices will be attained with less transport in the case of gravels than with sands.

The streams of the Black Hills are sup-

plied with gravel having very youthful maturity indices. Samples taken close to the foothills reveal youthful indices in terms of lithology but also mature indices in terms of rounding of some rock types. For example, gravel samples collected near the Dakota hogback average 35 per cent limestone and sandstone, a youthful index, but also contain well-rounded pebbles of quartz, a mature index. In Rapid Creek sample R-3 quartz pebbles 16–32 mm., after about 20 miles of transport, have an average roundness of 0.55. With a comparable distance of transport in Battle Creek, 1–1.414-mm. quartz sand has attained a roundness of only 0.31. It is evident from this comparison that a mature rounding index is more quickly reached in the case of large particles.

A gravel may be considered to have attained an index of old age when its component particles consist wholly of quartzose rocks (quartz, quartzite, chert) and those particles have reached their limit of rounding. The data suggest that mature indices are reached rather quickly in gravel deposits. In Rapid Creek 30 miles of transport have resulted in the loss of all but 2 or 3 per cent of the soft-rock components present in the gravels close to the Hills. In that distance the roundness of 16–32-mm. quartz pebbles has increased from 0.55 to 0.58. It is not possible, from the available data, to compute accurately the distance of transport necessary to reach the end of the maturity scale. However, from the chert ratio curves of figure 15 it is estimated that about 50 miles of transport would be required to remove essentially all but the quartzose rocks from the 16–32-mm. size class. The very slight increase in roundness from 0.55 to 0.58 in 30 miles suggests that the quartz pebbles have already approached an asymptotic roundness value in that distance.

It is apparent that the preceding discussion has only scratched the surface of the concept of "maturity indices." It is not sufficient to evaluate indices on the basis of any one size material, as has been done in the present work. The results obtained here suggest that a large range of sizes from many gravels and sands must be investigated from the maturity viewpoint before such a concept can become a practical tool for interpretation of clastic sediments. It is possible that further investigation may yield additional maturity indices. The degree of sorting of a sand or gravel or the standard deviation of roundness may furnish significant maturity criteria.

## SUMMARY AND CONCLUSIONS

The study of detritus transported by streams draining the Black Hills has furnished a number of conclusions important to an understanding of sedimentary processes. Some of the conclusions presented here verify field results of other workers, some have been hinted at by laboratory investigators, and still others are contradictory, or supplementary, to prevailing ideas of the effects of stream transport on sedimentary particles.

In terms of size analysis the following conclusions have been drawn for streams similar to those of the Black Hills.

1. The mean size of a gravel deposit decreases with distance of transport. The decrease in mean size results primarily from a decreasing competence of the stream as the gradient decreases. A direct relationship exists between stream gradient and mean size, irrespective of whether or not the stream has a graded profile.

Quantitatively, selective transport accounts for 75 per cent of the size decrease in Rapid Creek and 84 per cent in Battle

Creek. The remaining percentages are accounted for by abrasion and breakage.

2. The skewness of the size distribution of a gravel deposit decreases with distance of transport. This decrease is a direct result of a decrease in the mean size.

3. The standard deviation or sorting of a gravel deposit shows no systematic change with distance of transport.

4. Secondary modes in the size distributions of Rapid and Bear Butte creeks are attributed to sand infiltration subsequent to deposition of the coarser material. Similar secondary modes of Battle Creek are attributed to inclusion of extensive sand layers in the channel samples.

Analysis of roundness and sphericity data yields the following conclusions:

1. The rounding of sedimentary particles may be expressed by the differential equation

$$\frac{dP}{dS} = KS^{a}(P_L - P),$$

which states that the rate of change of roundness with distance is directly proportional to some power ($a$) of the distance ($S$) and to the difference between the roundness ($P$) at any point and a limiting value of roundness ($P_L$). The constants of this equation ($K$ and $a$) vary, depending on the composition and size of the particle, the rigor of transport to which it is subjected, and the size and composition of associated particles.

2. Sphericity change with distance shows a slight increase but is very erratic.

3. Quartz sand 1–1.414 mm. is definitely rounded in mountain streams of steep gradient. A 71 per cent increase in roundness is accomplished in the first 45 miles of transport.

4. Since both roundness and sphericity are related to size, they are related to each other. For 16–32-mm. limestone pebbles, a 12.3 per cent increase in sphericity is accompanied by a 380 per cent increase in roundness. Quartz sand 1–1.414 mm. yields a 119.1 per cent increase in roundness for a 4.7 per cent increase in sphericity.

Analysis of lithologic data leads to the following conclusions:

1. The initial lithologic frequency distribution of a gravel is directly related to the source area which furnishes detritus to a stream.

2. A short distance of stream transport removes most of the soft-rock types by abrasion and breakage. In a distance of 30 miles in Rapid Creek the percentage of limestone plus sandstone pebbles 16–32 mm. is reduced from 26.2 per cent to 2.6 per cent.

3. Loss of the softer rocks by abrasion and breakage during stream transport is a function of rigor of transport and size and composition of associated particles.

4. An order of resistance to abrasion and breakage is established for these streams. The rock types are listed in order of increasing resistance: sandstone, limestone-feldspar, pre-Cambrian metamorphics, quartz-quartzite, chert.

5. Mountain streams are very effective in reducing the feldspar content of coarse sands. A loss of 51.2 per cent in the feldspar content of 1–1.414-mm. sands resulted from a transport of 40 miles in Battle Creek. Transport of 150 miles in the Cheyenne River resulted in a 13.8 per cent decrease in feldspar content. This loss in feldspar can be attributed only to abrasion, breakage, and weathering during transport.

ACKNOWLEDGMENTS.—This work was made possible by the California Company Fellowship for research in sedimentology. The writer is also indebted to several individuals, whose

aid contributed to the carrying-out of the project. Mr. W. Schmidt, of the University of Chicago, helped design and construct sampling equipment for the field. Dr. George Otto, of the Armour Research Foundation, provided valuable information on techniques of field sampling. The writer's wife, Mrs. Donata Plumley, proved an indispensable assistant in the field work. Considerable aid in the mathematical treatment of pebble rounding was provided by Dr. Harold J. Plumley, of the Naval Ordnance Laboratory. Dr. L. Horberg, of the University of Chicago, suggested many helpful ideas on the interpretation of terraces. The writer is most indebted to Dr. F. J. Pettijohn, of the University of Chicago, whose helpful suggestions and criticisms have helped to carry the project through from its inception.

## REFERENCES CITED

Barrell, J. (1925) Marine and terrestrial conglomerates: Geol. Soc. America Bull. 36, pp. 279–342.

Crosby, W. O. (1882) Geology of the Black Hills of Dakota: Boston Soc. Nat. History Proc., vol. 23, pp. 516–517.

Dake, C. L. (1921) The problem of the St. Peter sandstone: Univ. Missouri School of Mines and Metallurgy Bull., Tech. Ser., vol. 6, chap. 2.

Darton, N. H. (1899–1900) Preliminary description of the geology and water resources of the southern half of the Black Hills and adjoining regions in South Dakota and Wyoming: U.S. Geol. Survey, 21st Ann. Rept., pt. 4, pp. 489–600.

——— (1909) Geology and water resources of the northern portion of the Black Hills: U.S. Geol. Survey Prof. Paper 65.

Fenneman, N. M. (1931) Physiography of western United States, New York, McGraw-Hill Book Co.

Fillman, L. (1929) Cenozoic history of the northern Black Hills: Univ. Iowa Studies in Nat. History, vol. 13, pp. 1–48.

Fraser, H. J. (1935) Experimental study of the porosity and permeability of clastic sediments: Jour. Geology, vol. 43, pp. 910–1010.

Graton, L. C., and Fraser, H. J. (1935) Systematic packing of spheres—with particular relation to porosity and permeability: Jour. Geology, vol. 43, pp. 785–909.

Ingerson, E., and Ramisch, J. L. (1942) Origin of shapes of quartz sand grains: Am. Mineralogist, vol. 27, pp. 595–606.

Kelsh, H. T. (1940) The slotted-templet method for controlling maps made from aerial photographs: U.S. Dept. of Agr. Misc. Pub. 404, pp. 1–29.

Krumbein, W. C. (1934) The probable error of sampling sediments for mechanical analysis: Am. Jour. Sci., vol. 27, pp. 204–214.

——— (1937) Sediments and exponential curves: Jour. Geology, vol. 45, pp. 577–601.

——— (1940) Flood gravel of San Gabriel Canyon, California: Geol. Soc. America Bull. 51, pp. 639–676.

——— (1941*a*) Measurement and geological significance of shape and roundness of sedimentary particles: Jour. Sedimentary Petrology, vol. 11, pp. 64–72.

——— (1941*b*) The effects of abrasion on the size, shape, and roundness of rock fragments: Jour. Geology, vol. 49, pp. 482–520.

——— (1942) Flood deposits of Arroyo Seco, Los Angeles County, California: Geol. Soc. America Bull. 53, pp. 1355–1402.

——— and Pettijohn, F. J. (1938) Manual of sedimentary petrography, New York, D. Appleton-Century.

——— and Tisdel, F. W. (1940) Size distribution of source rocks of sediments: Am. Jour. Sci., vol. 238, pp. 296–305.

Krynine, P. D. (1940) Petrology and genesis of the third Bradford Sand: Pennsylvania State Coll. Bull. 29.

Kurk, E. H. (1941) The problem of sampling heterogeneous sediments: Unpublished Master's dissertation, Dept. of Geology, University of Chicago, pp. 1–37.

MacClintock, P. (1936) A Pleistocene lake in the White River Valley: Am. Naturalist, vol. 70, pp. 346–360.

Mackie, W. (1896) The sands and sandstones of eastern Moray: Edinburgh Geol. Soc. Trans., vol. 7, pp. 148–172.

Meyerhoff, H. A., and Olmstead, E. W. (1937) Cenozoic leveling in the Black Hills (abstr.): Pan-Am. Geologist, vol. 68, p. 306.

Newton, H., and Jenney, W. (1880) Geology of the Black Hills of Dakota: U.S. Geol. and Geog. Survey.

Otto, G. H. (1933) Comparative tests of several methods of sampling heavy mineral concentrates: Jour. Sedimentary Petrology, vol. 3, pp. 30–39.

——— (1938) The sedimentation unit and its use in field sampling: Jour. Geology, vol. 46, pp. 569–582.

Perisho, E. C., and Visher, S. S. (1912) The geography, geology and biology of south-central South Dakota: South Dakota State Geol. and Biol. Survey Bull. 5, pp. 57–58.

Pettijohn, F. J., and Lundahl, A. C. (1943) Shape and roundness of Lake Erie beach sands: Jour. Sedimentary Petrology, vol. 13, pp. 69–78.

RUSSELL, R. D. (1937) Mineral composition of Mississippi River sands: Geol. Soc. America Bull. 48, pp. 1307–1348.

——— and TAYLOR, R. E. (1937) Roundness and shape of Mississippi River sands: Jour. Geology, vol. 45, pp. 225–267.

RUSSELL, W. L. (1929) Drainage alignment in the western Great Plains: Jour. Geology, vol. 37, pp. 249–255.

SARMIENTO, A. (1945) Experimental study of pebble abrasion: Unpublished Master's dissertation, Dept. of Geology, University of Chicago, pp. 1–59.

SHULITS, S. (1941) Rational equation of river-bed profile: Am. Geophys. Union Trans., 22d Ann. Meeting, pt. 2, pp. 622–630.

THIEL, G. A. (1940) The relative resistance to abrasion of mineral grains of sand size: Jour. Sedimentary Petrology, vol. 10, pp. 103–124.

TODD, J. E. (1900) A preliminary report on the geology of South Dakota: South Dakota Geol. Survey Bull. 1, pp. 120–121.

——— (1902) Hydrographic history of South Dakota: Geol. Soc. America Bull. 13, pp. 27–40.

——— (1923) Is the channel of the Missouri River through North Dakota of Tertiary origin?: Geol. Soc. America Bull. 34, pp. 491–493.

TWENHOFEL, W. H. (1945) The rounding of sand grains: Jour. Sedimentary Petrology, vol. 15, pp. 59–71.

UDDEN, J. A. (1914) Mechanical composition of clastic sediments: Geol. Soc. America Bull. 25, pp. 736–737.

WADELL, H. (1935) Volume, shape, and roundness of quartz particles: Jour. Geology, vol. 43, pp. 276–277.

WANLESS, H. R. (1923) The lithology and stratigraphy of the White River beds of South Dakota: Am. Philos. Soc. Proc., vol. 62, pp. 190–269.

WENTWORTH, W. C. (1919) A laboratory and field study of cobble abrasion: Jour. Geology, vol. 27, pp 507–522.

——— (1926) Methods of mechanical analysis of sediments: Univ. Iowa Studies in Nat. History, vol. 2, no. 2.

# THE UNIQUE ASSOCIATION OF THALLIUM AND RUBIDIUM IN MINERALS[1]

L. H. AHRENS
Massachusetts Institute of Technology

ABSTRACT

Spectrochemical analyses of various minerals has shown that "alkali-metal" thallium and rubidium are found only in potassium minerals and the cesium mineral, pollucite, and that in these minerals the Tl:Rb association is very close. Altogether, 167 specimens have been analyzed quantitatively, the selection comprising lepidolite, amazonite, hydrothermal pegmatitic microcline, primary pegmatitic microcline, zinnwaldite, biotite, muscovite, phlogopite, pollucite, rhodizite, and cesium beryl. The mean weight percentage ratio of $Rb_2O/Tl_2O$ was determined as 100; and the vast majority of the ratios fall within the limits of 35–300: the extreme limits are 10 and 650. A plot of log percentage of $Rb_2O$ versus log percentage of $Tl_2O$ produced a curve of unit slope over a thousand fold range of concentration that could be investigated. There thus appears to be no shift in the ratio Rb/Tl throughout the selective crystallization of minerals, and the ratio seems to be independent of type of host mineral; the ratio does, however, vary to some extent from area to area. The reasons for the close association of alkali-metal thallium and rubidium are that the radii of their ions are identical and that in certain pertinent respects their chemical properties are very similar.

With the possible exception of the pair Zr:Hf, which are very closely associated in minerals, alkali-metal thallium and rubidium are perhaps the most closely associated pair of elements in the earth's crust, and their association is all the more remarkable because thallium is a Group 3*b* element, whereas rubidium is an alkali metal (Group 1*a*).

The abundance of thallium in the earth's crust has been estimated as 0.0003 per cent by weight.

## INTRODUCTION

Soon after its discovery in 1861 by Sir William Crookes, chemists recognized the peculiar chemical characteristics of thallium which led Dumas, even in 1862, to dub this element the "ornithorhynchus" (the duckbill platypus) among the elements because of its paradoxical properties. It was soon found to be omnipresent, but in very small amounts, in several sulphide minerals and in a few as a major constituent; for example, crookesite $(Cu, Tl, Ag)_2Se$, lorandite ($TlAsS_2$); and hutchisonite $(Tl, Ag, Cu)_2S \cdot PbS \cdot 2As_2S_3$.

Thallium had also been found in some potassium minerals such as mica and feldspar, where, in considerable contrast to its distribution referred to above, it was thought that thallium entered potassium minerals by substituting for potassium. In this paper the geochemistry of thallium in its latter role, that is, simulating an alkali metal ("alkali-metal" thallium) is discussed. Quantitative abundance data are given on its distribution in most potassium minerals and the cesium mineral, pollucite; and, in particular, the association of thallium with rubidium has been investigated in these minerals.

Because thallium is a Group 3*b* element in the Periodic Table and, as will be shown later, because its association with rubidium is very close and unique, some pertinent aspects of the chemistry of this element will be examined as an aid to understanding its geochemistry.

## CHEMISTRY OF THALLIUM

Electronic configurations of each of the typical Group 3 elements and also the Group 3*b* elements are given in table 1.

Each of these elements has potentially three valence electrons (*s*- and *p*-orbits). In boron these three valence electrons are extremely tightly bound; but, with the

[1] Manuscript received March 9, 1948.

increase in the size of the atoms of the elements of higher atomic weight, the outer valence electrons become more loosely held, and variable valences are exhibited. In thallium the residual positive charge from the nucleus is weakest: the two 6*s* electrons tend to behave as an inert pair, and the completed shells plus the 6*s* electrons (gold core) are relatively stable, with the result that in contrast to other Group 3 elements, thallium is characteristically univalent. Lead, which follows thallium, shows a similar phenomenon, and divalent stable lead salts are common. The formation of a stable mation of a sparingly soluble sulphide and chloride. Its sulphate and carbonate and most other thallium compounds are very similar to those of the alkali metals.

Since $Tl^+$ and $Rb^+$ have identical radii and as the chemical properties of these two elements are very similar in many important respects, thallium and rubidium should be very closely associated in most minerals, that is, in those minerals in which precipitation of thallous sulphide or chloride is not likely to occur. This investigation has been confined in the main, therefore, to silicate minerals of igneous origin.

TABLE 1

ELECTRONIC CONFIGURATIONS

| Element | Atomic Number | K | L | | M | | | N | | | | O | | | P | | | (Shell) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | 1s | 2s | 2p | 3s | 3p | 3d | 4s | 4p | 4d | 4f | 5s | 5p | 5d | 6s | 6p | 6d | (Orbit) |
| B | 5 | 2 | 2 | 1 | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... |
| Al | 13 | 2 | 2 | 6 | 2 | 1 | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... |
| Ga | 31 | 2 | 2 | 6 | 2 | 6 | 10 | 2 | 1 | ... | ... | ... | ... | ... | ... | ... | ... | ... |
| In | 49 | 2 | 2 | 6 | 2 | 6 | 10 | 2 | 6 | 10 | ... | 2 | 1 | ... | ... | ... | ... | ... |
| Tl | 81 | 2 | 2 | 6 | 2 | 6 | 10 | 2 | 6 | 10 | 14 | 2 | 6 | 10 | 2 | 1 | ... | ... |

monovalent thallous ion, in place of a trivalent one, causes a marked expansion in size. The radius of $Tl^+ = 1.49$ A, which radius equals that of $Rb^+$. Next to Cs (radius = 1.69 A) these ions are the largest cations. As the ionic radii of $Tl^+$ and $Rb^+$ are identical, the facility with which the two ions enter lattice hosts should be very similar.

In certain respects which are of geochemical significance, thallous thallium exhibits chemical properties almost identical with those of the ions of the alkali metals of higher atomic weight. $Tl^+$ is a typical large-sized cation and in aqueous solution forms a soluble hydroxide, TlOH, which is a caustic alkali of similar strength to NaOH, KOH, etc. Thallium differs from the alkali metals in the for-

It may be recalled that, although some other pairs of elements have reasonably close chemical properties and have ionic radii which are listed as identical or very nearly so, their geochemical association is not unusually close. The pair, nickel: magnesium, is an example, but it is known that, whereas the $Mg^{++}$-anion bond is essentially ionic, the $Ni^{++}$-anion bond is partly covalent, thereby causing an apparent decrease in the size of $Ni^{++}$. The pair, zirconium:hafnium, have ions of identical size and show the same type of bonding, and hence these two elements are very closely associated in minerals. In silicate minerals, the $Rb^+$-anion bond and $Tl^+$-anion bond are undoubtedly ionic, and one may regard the effective radii of $Rb^+$ and $Tl^+$ as equal.

### SELECTION OF MINERALS FOR ANALYSIS

A preliminary qualitative spectrochemical survey of several minerals corroborated earlier evidence of the presence of thallium and rubidium in all potassium minerals and also in the cesium mineral, pollucite. In sodium-rich minerals, only occasionally could slight traces of rubidium and thallium be detected. The selection therefore comprises chiefly potassium minerals of various types, some specimens of pollucite, a few specimens of sodium minerals, and a few other types. Reference to the analytical data given in table 2 provides details of the minerals analyzed.

### ANALYTICAL PROCEDURE

All analyses of thallium and rubidium were made spectrochemically. A description of the spectrochemical procedure employed has been given in an earlier paper by the author (1945) on the geochemical association of thallium and rubidium.

Most analyses were made on the large quartz spectrograph of the Government Metallurgical Laboratory, University of the Witwatersrand, Johannesburg, and about twenty-five specimens were analyzed with a large-grating instrument of the Department of Geology, Massachusetts Institute of Technology. The large dispersion (2.5 A/mm) of the latter instrument was necessary for analyzing biotite, in the analysis of which iron caused excessive interference when quartz optics were used.

### ANALYTICAL RESULTS

The quantities of thallium and of rubidium found in the mineral specimens, together with the respective $Rb_2O/Tl_2O$, ratios (by weight percentage) are given in table 2. In addition to these analyses, Adamson (1942) has determined thallium and rubidium in a specimen of hydrothermal microcline from Varuträsk, Sweden, and found 2.1 per cent $Rb_2O$ and 0.02 per cent $Tl_2O$; ratio $RB_2O/Tl_2O = 100$.

### DISCUSSION OF TABLE 2

#### VARIATION OF THE PERCENTAGE RATIO $Rb_2O/Tl_2O$ IN GENERAL

In all, 167 ratios could be determined. The change in ratio is best shown graphically, and figure 1 shows the percentage of $Rb_2O$ plotted against the percentage of $Tl_2O$, using log co-ordinates. A straight line of unit slope fits the points most satisfactorily, and, throughout the concentration range of about one thousand which could be investigated, there is no apparent change in the ratio, the mean value being, according to the graph, very nearly 100; the corresponding atomic ratio, Rb/Tl, is equivalent to 225. The value of this ratio seems to be independent of the host mineral but appears to vary from area to area and sometimes with a particular phase of mineralization, about which more will be said later.

As a reasonably large number of analyses were available, an attempt was made to draw a probability (known also as "normal" or "Gaussian") curve. If an accurate curve of this type can be drawn, a determination may be made of the frequency with which different values of the ratio $Rb_2O/Tl_2O$ deviate from the mean, the value of which may be obtained from this curve. Figure 2 is a plot of frequency (ordinate) with which the ratio $Rb_2O/Tl_2O$ falls within a definite interval in the ratio versus the logarithm (to base 10) of the ratio, plotted at each interval. An interval factor of 1.4 was found convenient, and each interval used and the frequency with which the ratio fell within the limits of each interval are tabulated in table 3; from these data figure 2 was plotted.

## TABLE 2

| No. | Locality | Per Cent $Rb_2O$ | Per Cent $Tl_2O$ | $Rb_2O/Tl_2O$ | Per Cent $K_2O$ |
|---|---|---|---|---|---|
| | A. Potash Feldspar | | | | |
| 1. *Pegmatitic** | | | | | |
| 1 | Mica Siding, E. Transvaal, S. Africa | 0.022 | 0.00038 | 60 | ...... |
| 2 | Mica Siding, E. Transvaal, S. Africa | 0.024 | 0.00037 | 65 | ...... |
| 3 | Mica Siding, E. Transvaal, S. Africa | 0.018 | 0.00027 | 70 | ...... |
| 4 | W. of Kaleka, Krug. Nat. Pk., S. Africa | 0.013 | 0.00017 | 75 | ...... |
| 5 | N. of Letaba River causeway, Krug. Nat. Pk., S. Africa | 0.0075 | 0.00008 | 90 | ...... |
| 6 | N. of Letaba River, etc. | 0.010 | 0.00012 | 80 | ...... |
| 7 | S. of Loole Kop, E. Transvaal | 0.005 | 0.00005 | 100 | ...... |
| 8 | Major, Klein Letaba area, E. Transvaal | 0.011 | 0.00010 | 110 | ...... |
| 9 | Olifantsrivierspoort, E. Transvaal | 0.0045 | 0.000055 | 80 | ...... |
| 10 | Cotton Kop, E. Transvaal | 0.022 | 0.00005 | 440 | ...... |
| 11 | Loole Kop, E. Transvaal | 0.024 | 0.00006 | 400 | ...... |
| 12 | Loole Kop, E. Transvaal | 0.020 | 0.00005 | 400 | ...... |
| 13 | Loole Kop, E. Transvaal | 0.024 | 0.00009 | 270 | ...... |
| 14 | Palabora, E. Transvaal | 0.018 | 0.00005 | 360 | ...... |
| 15 | Loole Kop, E. Transvaal | 0.013 | 0.00005 | 260 | ...... |
| 16 | W. of Loole Kop, E. Transvaal | 0.033 | 0.00005 | 650 | ...... |
| 17 | Palabora, E. Transvaal | 0.025 | 0.00005 | 500 | ...... |
| 18 | Palabora, E. Transvaal | 0.015 | 0.00004 | 400 | ...... |
| 19 | S.W. of Shangaan, E. Transvaal | 0.015 | 0.00006 | 250 | ...... |
| 20 | S.W. of Shangaan, E. Transvaal | 0.020 | 0.00005 | 400 | ...... |
| 21 | S.W. of Shangaan, E. Transvaal | 0.025 | 0.00005 | 500 | ...... |
| 22 | S.W. of Shangaan, E. Transvaal | 0.033 | 0.00010 | 330 | ...... |
| 23 | Letaba area, E. Transvaal | 0.011 | 0.00004 | 300 | ...... |
| 24 | Johannesburg, Transvaal | 0.077 | 0.00052 | 150 | ...... |
| 25 | Johannesburg, Transvaal | 0.088 | 0.00062 | 140 | ...... |
| 26 | Johannesburg, Transvaal | 0.078 | 0.00060 | 130 | ...... |
| 27 | Johannesburg, Transvaal | 0.060 | 0.00050 | 120 | ...... |
| 28 | Johannesburg, Transvaal | 0.058 | 0.00058 | 100 | ...... |
| 29 | Johannesburg, Transvaal | 0.28 | 0.0021 | 130 | ...... |
| 30 | Johannesburg, Transvaal | 0.066 | 0.00060 | 110 | ...... |
| 31 | Johannesburg, Transvaal | 0.040 | 0.00060 | 70 | ...... |
| 32 | Mulder's Drift, near Johannesburg | 0.033 | 0.00034 | 100 | ...... |
| 33 | Mulder's Drift, near Johannesburg | 0.035 | 0.00033 | 110 | ...... |
| 34 | Rhenoster Spruit, Waterburg, Transvaal | 0.028 | 0.00016 | 180 | ...... |
| 35 | Mutue Fides, Transvaal | 0.030 | 0.00010 | 300 | ...... |
| 36 | Leeuwpoort Tin Mine, Transvaal | 0.018 | 0.00010 | 180 | ...... |
| 37 | Messina, Transvaal | 0.014 | 0.00010 | 140 | ...... |
| 38 | S. of Port Shepstone, Natal, S. Africa | 0.010 | 0.00016 | 60 | ...... |
| 39 | Near Umgeni Dam, Natal, S. Africa | 0.015 | 0.00005 | 300 | ...... |
| 40 | Oribi Gorge, Natal, S. Africa | 0.019 | 0.00012 | 160 | ...... |
| 41 | Valley of 1,000 Hills, Natal, S. Africa | 0.009 | 0.00025 | 40 | ...... |
| 42 | Illovo River, Mid-Illovo, Natal, S. Africa | 0.020 | 0.00032 | 70 | ...... |
| 43 | Illovo River, Mid-Illovo, Natal, S. Africa | 0.0084 | 0.00025 | 35 | ...... |
| 44 | Illovo River, Mid-Illovo, Natal, S. Africa | 0.005 | 0.00013 | 40 | ...... |
| 45 | Illovo River, Mid-Illovo, Natal, S. Africa | 0.006 | 0.00012 | 50 | ...... |
| 46 | Gordonia, NW. Cape Province, S. Africa | 0.008 | 0.00016 | 55 | ...... |
| 47 | Henkries Valley, Namaqualand, S. Africa | 0.037 | 0.00031 | 120 | ...... |
| 48 | Uranoop River area, Namaqualand, S. Africa | 0.060 | 0.00072 | 80 | ...... |
| 49 | Uranoop River area, Namaqualand, S. Africa | 0.043 | 0.00044 | 100 | ...... |
| 50 | Uranoop River area, Namaqualand, S. Africa | 0.050 | 0.00069 | 70 | ...... |
| 51 | Uranoop River area, Namaqualand, S. Africa | 0.012 | 0.00014 | 90 | ...... |
| 52 | Uranoop River area, Namaqualand, S. Africa | 0.017 | 0.00016 | 100 | ...... |
| 53 | Uranoop River area, Namaqualand, S. Africa | 0.056 | 0.00056 | 100 | ...... |
| 54 | Uranoop River area, Namaqualand, S. Africa | 0.047 | 0.00050 | 90 | ...... |
| 55 | Bantry Bay, Cape Province, S. Africa | 0.019 | 0.00033 | 60 | ...... |

* Chiefly microcline and perthite.

TABLE 2—*Continued*

| No. | Locality | Per Cent $Rb_2O$ | Per Cent $Tl_2O$ | $Rb_2O/Tl_2O$ | Per Cent $K_2O$ |
|---|---|---|---|---|---|
| | A. Potash Feldspar—*Continued* | | | | |
| 2. *From granite* | | | | | |
| 56...... | Loole Kop, E. Transvaal, S. Africa | 0.006 | 0.00008 | 75 | ...... |
| 57...... | N.E. of Loole Kop, Transvaal, S. Africa | 0.009 | 0.00011 | 80 | ...... |
| 58...... | Mashishimala Hills, Transvaal, S. Africa | 0.007 | 0.00007 | 100 | ...... |
| 3. *Hydrothermal microcline from pegmatite*† | | | | | |
| 59...... | Harding Mine, Dixon, New Mexico | 0.21 | 0.005 | 40 | ...... |
| 60...... | Harding Mine, Dixon, New Mexico | 0.68 | 0.019 | 35 | ...... |
| 61...... | Harding Mine, Dixon, New Mexico | 0.41 | 0.013 | 30 | ...... |
| 4. *Amazonite* | | | | | |
| 62...... | Honeydew, Zoutpansberg, Transvaal | 0.82 | 0.0052 | 160 | ...... |
| 63...... | Klein Spitzkop, SW. Africa | 0.15 | 0.0010 | 150 | ...... |
| 64...... | Klein Spitzkop, SW. Africa | 0.13 | 0.0010 | 130 | ...... |
| 65...... | Klein Spitzkop, SW. Africa | 0.20 | 0.0012 | 170 | ...... |
| 66...... | Klein Spitzkop, SW. Africa | 0.17 | 0.0010 | 170 | ...... |
| 67...... | Maltahöhe, SW. Africa | 0.33 | 0.0031 | 110 | ...... |
| 68...... | Otjiwarongo, SW. Africa | 0.44 | 0.0018 | 240 | ...... |
| 69...... | Xamchab, Namaqualand, S. Africa | 0.22 | 0.0016 | 140 | ...... |
| 70...... | Gordonia, NW. Cape Province | 0.25 | 0.0020 | 120 | ...... |
| 71...... | Madagascar | 1.30 | 0.0036 | 360 | ...... |
| 72...... | Labrador | 0.13 | 0.0010 | 130 | ...... |
| 73...... | Norway | 0.40 | 0.0080 | 50 | ...... |
| 74...... | Ontario, Canada | 0.33 | 0.0010 | 330 | ...... |
| 75...... | Pike's Peak, Colorado | 1.30 | 0.0052 | 260 | ...... |
| 76...... | Florissant, Colorado | 0.19 | 0.00062 | 300 | ...... |
| 77...... | ? | 0.07 | 0.00062 | 110 | ...... |
| 5. *Unclassified* | | | | | |
| 78...... | Sweden | 0.025 | 0.00041 | 60 | ...... |
| 79...... | New Jersey | 0.013 | 0.00034 | 35 | ...... |
| 80...... | Baveno, Italy (orthoclase) | 0.055 | 0.00062 | 90 | ...... |
| 81...... | Wehr-Eifel, Germany | 0.010 | 0.00003? | 350? | ...... |
| 82...... | St. Gothard, Switzerland | 0.026 | 0.00062 | 40 | ...... |
| 83...... | ? | 0.014 | 0.00013 | 110 | ...... |
| 84...... | ? | 0.015 | 0.00034 | 45 | ...... |
| 85...... | ? | 0.023 | 0.00010 | 230 | ...... |
| | B. Plagioclase, Chiefly Albite‡ | | | | |
| 86...... | Bandolier Kop, Transvaal | 0.003? | .......... | .......... | 0.23 |
| 87...... | Steinkopf Reserve, Namaqualand | .......... | .......... | .......... | 0.10 |
| 88...... | Uranoop River, Namaqualand | .......... | .......... | .......... | 0.10 |
| 89...... | Uranoop River, Namaqualand | .......... | .......... | .......... | 0.10 |
| 90...... | Uranoop River, Namaqualand | .......... | .......... | .......... | 0.22 |
| 91...... | Uranoop River, Namaqualand | .......... | .......... | .......... | 0.20 |
| 92...... | Uranoop River, Namaqualand | .......... | .......... | .......... | 0.10 |
| 93...... | Selati River, NE. Transvaal | .......... | 0.00004 | .......... | 0.20 |
| 94...... | Warmbad, SW. Africa | 0.012 | 0.0003 | 40 | 0.28 |
| 95...... | Karibib, SW. Africa | 0.002? | 0.00003? | .......... | 0.10 |
| 96...... | Sea Point, Cape Province | .......... | .......... | .......... | 0.80 |
| 97...... | Southern Rhodesia | 0.002? | .......... | .......... | 0.40 |
| 98...... | Southern Rhodesia | .......... | .......... | .......... | 1.0 |

† These three specimens have been listed as hydrothermal. According to A. Montgomery (private communication), who collected the specimens, the latter two specimens, one of which is very pale green and similar to amazonite, are of late formation and are believed to be hydrothermal. The first specimen is considered to be of earlier formation and could probably be better included with the common (primary) microcline group.

‡ To show the essentially sodic nature of these feldspars, the $K_2O$ contents (determined spectrochemically) are also included.

TABLE 2—*Continued*

| No. | Locality | Per Cent $Rb_2O$ | Per Cent $Tl_2O$ | $Rb_2O/Tl_2O$ | Per Cent $K_2O$ |
|---|---|---|---|---|---|
| | **C. Lepidolite** | | | | |
| 99 | Okongava Ost 72, Karibib, SW. Africa | 1.80 | 0.023 | 70 | ...... |
| 100 | Okongava Ost 72, Karibib, SW. Africa | 1.30 | 0.016 | 80 | ...... |
| 101 | Albrecht's Höhe, SW. Africa | 0.88 | 0.017 | 50 | ...... |
| 102 | Omaruru, SW. Africa | 2.40 | 0.013 | 180 | ...... |
| 103 | Otjimboyo, SW. Africa | 0.90 | 0.016 | 55 | ...... |
| 104 | Arandis, SW. Africa | 0.47 | 0.010 | 50 | ...... |
| 105 | Warmbad, SW. Africa | 1.60 | 0.016 | 100 | ...... |
| 106 | Jackalswater, Namaqualand | 1.80 | 0.0052 | 350 | ...... |
| 107 | Jackalswater, Namaqualand | 2.00 | 0.0060 | 350 | ...... |
| 108 | Mozambique, E. Africa | 1.90 | 0.019 | 100 | ...... |
| 109 | Harding Mine, Dixon, New Mexico | | | | |
| 110 | Harding Mine { low lithia content | 0.41 | 0.011 | 35 | ...... |
| | Harding Mine { high lithia content | 1.0 | 0.022 | 45 | ...... |
| 111 | Portland, Connecticut | 0.78 | 0.010 | 80 | ...... |
| 112 | Strickland Quarry, Connecticut | 0.55 | 0.0074 | 75 | ...... |
| 113 | Pala, California | 2.00 | 0.021 | 100 | ...... |
| 114 | Pala, California | 2.00 | 0.018 | 110 | ...... |
| 115 | Norway, Maine | 1.70 | 0.016 | 110 | ...... |
| 116 | Auburn, Maine | 0.87 | 0.016 | 55 | ...... |
| 117 | Black Hills, South Dakota | 1.30 | 0.0062 | 210 | ...... |
| 118 | Black Hills, South Dakota | 1.30 | 0.009 | 150 | ...... |
| 119 | Brown Derby, Colorado | 2.80 | 0.024 | 120 | ...... |
| 120 | Brown Derby, Colorado | 2.60 | 0.028 | 90 | ...... |
| 121 | Copper Mt., N. of Bonneville, Wyoming | 1.40 | 0.010 | 140 | ...... |
| 122 | Along Winnipeg River, SE. Manitoba | 1.60 | 0.017 | 100 | ...... |
| 123 | Lake of the Woods, SE. Manitoba | 1.20 | 0.013 | 90 | ...... |
| 124 | Silver Leaf Mine, Winnipeg River, SE. Manitoba | 2.90 | 0.012 | 240 | ...... |
| 125 | Lipowka, Urals | 1.50 | 0.010 | 150 | ...... |
| 126 | Moravia | 1.45 | 0.013 | 110 | ...... |
| 127 | ? | 1.70 | 0.019 | 90 | ...... |
| | **D. Biotite (Lithium-rich from Pegmatite)** | | | | |
| 128 | Strickland Quarry, Connecticut | 0.47 | 0.0084 | 55 | ...... |
| 129 | King's Mt., North Carolina | 1.40 | 0.040 | 35 | ...... |
| | **E. Zinnwaldite** | | | | |
| 130 | Saxony, Germany | 0.66 | 0.0075 | 90 | ...... |
| 131 | Erzgebirge, Germany | 0.35 | 0.0065 | 55 | ...... |
| 132 | Virginia | 0.45 | 0.0074 | 60 | ...... |
| | **F. Muscovite** | | | | |
| 133 | Mica Siding, E. Transvaal | 0.066 | 0.0010 | 70 | ...... |
| 134 | Mica Siding, E. Transvaal | .......... | .......... | .......... | ...... |
| 135 | Olifants River, E. Transvaal | 0.050 | 0.0012 | 40 | ...... |
| 136 | Game Reserve, E. Transvaal | 0.15 | 0.0028 | 55 | ...... |
| 137 | Selati River, E. Transvaal | .......... | .......... | .......... | ...... |
| 138 | Near Leydsdorp, E. Transvaal | .......... | .......... | .......... | ...... |
| 139 | Olifants River, E. Transvaal | 0.027 | 0.00034 | 80 | ...... |
| 140 | South of Usakos, SW. Africa | 0.044 | 0.00062 | 70 | ...... |

| No. | Locality | Per Cent $Rb_2O$ | Per Cent $Tl_2O$ | $Rb_2O/Tl_2O$ | Per Cent $K_2O$ |
|---|---|---|---|---|---|
| | F. Muscovite—*Continued* | | | | |
| 141 | South of Usakos, SW. Africa | 0.049 | 0.00070 | 70 | ...... |
| 142 | Uranoop River area, Namaqualand | 0.096 | 0.0030 | 30 | ...... |
| 143 | Uranoop River area, Namaqualand | 0.045 | 0.0011 | 40 | ...... |
| 144 | Uranoop River area, Namaqualand | 0.14 | 0.0035 | 40 | ...... |
| 145 | Uranoop River area, Namaqualand | 0.10 | 0.0030 | 30 | ...... |
| 146 | Uranoop River area, Namaqualand | 0.053 | 0.0014 | 40 | ...... |
| 147 | Uranoop River area, Namaqualand | 0.10 | 0.0025 | 40 | ...... |
| 148 | Uranoop River area, Namaqualand | 0.087 | 0.0023 | 40 | ...... |
| 149 | Uranoop River area, Namaqualand | 0.010 | 0.00021 | 50 | ...... |
| 150 | Uranoop River area, Namaqualand | 0.011 | 0.00031 | 35 | ...... |
| 151 | Uranoop River area, Namaqualand | 0.063 | 0.0017 | 35 | ...... |
| 152 | Uranoop River area, Namaqualand | 0.045 | 0.0010 | 45 | ...... |
| 153 | Uranoop River area, Namaqualand | 0.066 | 0.0019 | 35 | ...... |
| 154 | Steinkopf Reserve, Namaqualand | 0.46 | 0.0024 | 190 | ...... |
| 155 | Near Steinkopf, Namaqualand | 0.64 | 0.0060 | 110 | ...... |
| 156 | NW. Cape Province | 0.008 | 0.0002 | 40 | ...... |
| 157 | NW. Cape Province | 0.11 | 0.0025 | 40 | ...... |
| 158 | Illovo River, Mid-Illovo, Natal | 0.042 | 0.0010 | 40 | ...... |
| 159 | Sea Point, Cape Province | 0.17 | 0.0038 | 45 | ...... |
| 160 | Bantry Bay, Cape Province | 0.02 | 0.00025 | 80 | ...... |
| 161 | George, Cape Province | 0.022 | 0.00037 | 60 | ...... |
| 162 | Miami, Southern Rhodesia | 0.028 | 0.00062 | 45 | ...... |
| 163 | Southern Rhodesia | 0.025 | 0.00031 | 60 | ...... |
| 164 | Salisbury, Southern Rhodesia | 0.30 | 0.0033 | 90 | ...... |
| 165 | Harding Mine, Dixon, New Mexico (rose muscovite) | 0.17 | 0.015 | 12 | ...... |
| 166 | North Carolina | 0.078 | 0.0023 | 35 | ...... |
| 167 | Pennsylvania | 0.020 | 0.00057 | 35 | ...... |
| 168 | Tyrol | ...... | ...... | ...... | ...... |
| 169 | Italy | 0.01 | 0.00020 | 50 | ...... |
| | G. Phlogopite and Vermiculitized Phlogopite | | | | |
| 170 | Graphic granite vein in gabbroidal rock, Vlakfontein, Transvaal | 0.035 | 0.00046 | 80 | ...... |
| 171 | Dunite pipe, Mooihoek, Transvaal | 0.042 | 0.00031 | 140 | ...... |
| 172 | Palabora, Transvaal (in syenite) | 0.072 | 0.00020 | 360 | ...... |
| 173 | Palabora, Transvaal (in syenite) | 0.055 | 0.00012 | 420 | ...... |
| 174 | Palabora, Transvaal (in syenite) | 0.050 | <0.00010 | ...... | ...... |
| 175 | Palabora, Transvaal (in syenite) | 0.036 | <0.00010 | ...... | ...... |
| 176 | Palabora, Transvaal (in syenite) | 0.012 | ...... | ...... | ...... |
| 177 | Palabora, Transvaal (in syenite) | 0.005–0.01 | ...... | ...... | ...... |
| 178 | Palabora, Transvaal (in syenite) | ...... | ...... | ...... | ...... |
| 179 | Antwerp, New York | 0.029 | 0.00029 | 100 | ...... |
| 180 | Ontario, Canada | 0.012 | 0.00030 | 40 | ...... |
| | H. Pollucite | | | | |
| 181 | Okongava Ost 72, Karibib, SW. Africa | 0.54 | 0.011 | 50 | ...... |
| 182 | Tin Mt., Black Hills, S. Dakota | 0.25 | 0.0019 | 130 | ...... |
| 183 | Greenwood, Maine | 0.68 | 0.0019 | 350 | ...... |
| 184 | Norway, Maine | 0.23 | 0.0013 | 180 | ...... |
| 185 | Varuträsk, Sweden | 0.37 | 0.008 | 45 | ...... |

TABLE 2—*Continued*

| No. | Locality | Per Cent $Rb_2O$ | Per Cent $Tl_2O$ | $Rb_2O/Tl_2O$ | Per Cent $K_2O$ |
|---|---|---|---|---|---|
| | I. Sundry Minerals | | | | |
| 186 | Beryl: Steinkopf Reserve, Namaqualand | ........ | ........ | ........ | ...... |
| 187 | Beryl: Manjak, Madagascar (contains 0.5% $Cs_2O$) | 0.30 | 0.004 | 80 | ...... |
| 188 | Rhodizite: Manjak, Madagascar (contains 1.5% $Cs_2O$) | 1.00 | 0.01 | 100 | ...... |
| 189 | Petalite: Karibib, SW. Africa | ........ | ........ | ........ | ...... |
| 190 | Amblygonite: Karibib, SW. Africa | ........ | ........ | ........ | ...... |
| 191 | Paragonite: Carinthia | ........ | ........ | ........ | ...... |

The equation for a normal curve is

$$y = k\,e^{-h^2x^2}\,,$$

where $k$ is the height of the curve when $x$, the deviation from the mean, is equal to zero. The ordinate($y$) is frequency. When $x = 0$, $y = k$. The symbol $h$ is a constant which determines the slope of the curve for a given $k$. Using values of $k = 29$ and $h = 0.062$, which are obtained by trial, a normal curve may be drawn which closely fits the curve shown in figure 2. One may conclude, therefore, that the curve in figure 2, which was not drawn from calculation, shows a near-normal Gaussian distribution of events about a mean.

With the aid of this curve, the following data are given on the $Rb_2O/Tl_2O$ ratio in igneous silicate minerals throughout the earth's crust as a whole (the mean ratio, $Rb_2O/Tl_2O$ is 90):

$\frac{1}{2}$ all ratios lie within $\frac{2}{3}$–$\frac{3}{2}$ of the mean, namely, within 60–135,
$\frac{2}{3}$ all ratios lie within $\frac{1}{2}$–$\frac{2}{1}$ of the mean, namely, within 45–180,
$\frac{4}{5}$ all ratios lie within $\frac{1}{3}$–$\frac{3}{1}$ of the mean, namely, within 30–270,
$\frac{24}{25}$ all ratios lie within $\frac{1}{5}$–$\frac{5}{1}$ of the mean, namely, within 20–450,
$\frac{99}{100}$ all ratios lie within $\frac{1}{10}$–$\frac{10}{1}$ of the mean, namely, within 10–900.

VARIATION OF THE RATIO $Rb_2O/Tl_2O$ IN SPECIFIC AREAS

Reference to the analyses given in table 2 shows clearly that different areas are characterized by definite ratios. In certain areas several mineral specimens have been analyzed; the value of the $Rb_2O/Tl_2O$ ratio will be discussed in some of these areas.

*"Old Granite"*[2] *area of the Olifants River.*—Ten specimens of feldspar and seven of muscovite (nos. 56, 1–9, 133–139 in table 2) were analyzed from this area. A few of the specimens are of doubtful association, that is, it is not quite clear whether they are genetically related to the "Old Granite" or to the syenites of the Palabora Complex. Because minerals that are definitely related to the "Old Granite" have typically low ratios, whereas those related to the Palabora syenites have characteristically high ratios, in cases of doubtful association where the ratios are low they are included with the "Old Granite," whereas where they are high they have been included with the Palabora syenite (Ahrens, 1945).

[2] In South Africa, the ancient granite-gneiss complexes are commonly referred to as the "Old Granite."

Almost within the limits of experimental error, the ratio $Rb_2O/Tl_2O$ is constant, irrespective of the mineral type.

*Uranoop River area, Namaqualand.*—Twenty-four specimens (twelve of muscovite, seven of microcline, and five of albite) from the area between Uranoop Trig beacon I and the Uranoop River, Namaqualand, were very kindly collected by Dr. John de Villiers of the Geological Survey, Union of South Africa. In each instance feldspar and mica are associated in the same pegmatite, but each feldspar-mica pair comes from a different pegmatite. For convenience of discussion, the analyses of these specimens are given again in table 4.

Some interesting conclusions may be drawn from these results:

1. The ratio $Rb_2O/Tl_2O$ is constant for all micas; indeed, this ratio is so constant that a plot of log $Rb_2O$ versus log $Tl_2O$ gives a straight line of unit slope, from which curve the plotted points deviate very slightly. In this example, as in others discussed under this subheading, the ratio $Rb_2O/Tl_2O$ approaches a constancy of a ratio of isotopes, when some allowance is made for the experimental error.

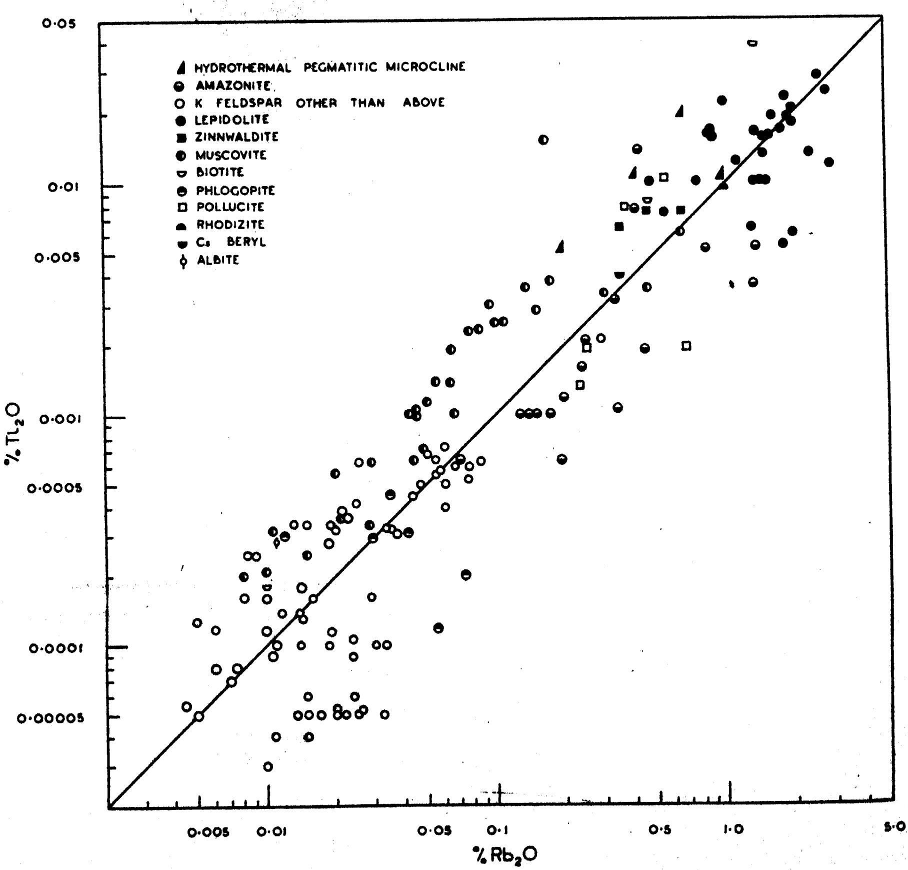



FIG. 1.—The relationship between thallium and rubidium in 167 mineral specimens of various types

2. The ratio $Rb_2O/Tl_2O$ is constant for all microcline specimens but differs from the value of the mica ratio.

3. Those mica specimens associated with albite are usually richer in rubidium and thallium than are those associated with microcline.

4. The rubidium contents of microcline and associated mica are approximately the same in each instance.

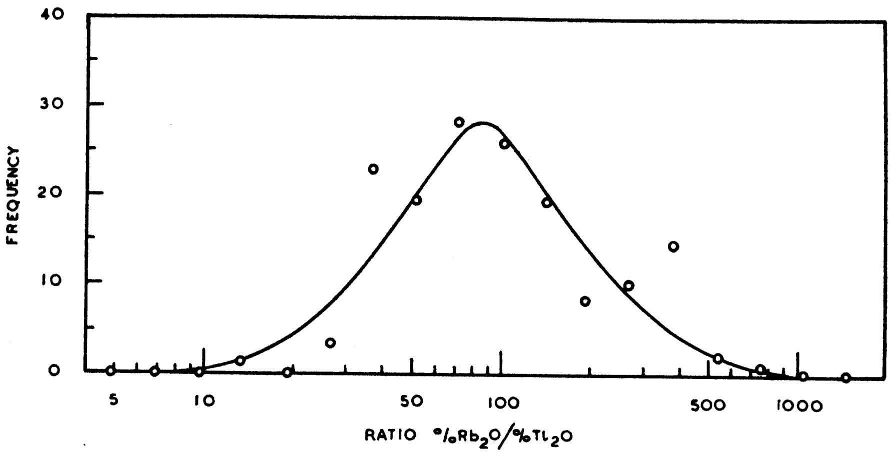



FIG. 2.—Normal (probability) curve, showing distribution of the variation of the ratio per cent $Rb_2O$: per cent $Tl_2O$ about the mean.

5. Neither thallium nor rubidium could be detected in any of the five specimens of albite.

Further discussion of the $Rb_2O/Tl_2O$ ratio in this area, as well as in the Palabora and the Usakos-Karibib areas, has been made by Ahrens (1945).

#### THE ABUNDANCE OF THALLIUM AND OF RUBIDIUM IN DIFFERENT MINERALS

A few comments will be made on the abundance of thallium and of rubidium in some of the minerals listed in table 2.

*Lepidolite.*—All specimens are characteristically enriched in thallium and rubidium, and lepidolite may be regarded as the mineral richest in alkali-metal thallium and rubidium. Mean values are 0.015 per cent $Tl_2O$ and 1.5 per cent $Rb_2O$.

*Amazonite and hydrothermal pegmatitic microcline.*—In contrast to microcline from granite and the common primary-phase microcline from pegmatite, amazonite (considered to be hydrothermal) and hydrothermal pegmatitic microcline appear to be markedly enriched in thal-

TABLE 3

| Interval | Frequency |
|---|---|
| 5.6– 7.9 | 0 |
| 8.9– 11.0 | 0 |
| 12.0– 15 | 1 |
| 16 – 22 | 0 |
| 23 – 30 | 3 |
| 31 – 42 | 23 |
| 43 – 59 | 19 |
| 60 – 83 | 28 |
| 84 – 116 | 26 |
| 117 – 162 | 19 |
| 163 – 223 | 8 |
| 224 – 317 | 10 |
| 318 – 445 | 14 |
| 446 – 620 | 2 |
| 621 – 870 | 1 |
| 871 –1200 | 0 |
| 1201 –1700 | 0 |
| 1701 –2400 | 0 |

lium and rubidium. Goldschmidt *et al.* (1934) have referred to the enrichment of rubidium in amazonite, although analyses by Erämetsä *et al.* (1941) of various specimens of microcline, including amazonite, do not support this characteristic enrichment of rubidium.

TABLE 4

| Mineral | Per Cent $Rb_2O$ | Per Cent $Tl_2O$ | $Rb_2O/Tl_2O$ |
|---|---|---|---|
| Microcline | 0.06 | 0.00072 | 80 |
| Muscovite | 0.096 | 0.0030 | 30 |
| Microcline | 0.043 | 0.00044 | 100 |
| Muscovite | 0.045 | 0.0011 | 40 |
| Microcline | 0.050 | 0.00069 | 70 |
| Muscovite | 0.053 | 0.0014 | 40 |
| Microcline | 0.012 | 0.00014 | 90 |
| Muscovite | 0.010 | 0.00021 | 50 |
| Microcline | 0.017 | 0.00016 | 100 |
| Muscovite | 0.011 | 0.0031 | 40 |
| Microcline | 0.056 | 0.00056 | 100 |
| Muscovite | 0.063 | 0.00017 | 35 |
| Microcline | 0.047 | 0.00050 | 90 |
| Muscovite | 0.045 | 0.0010 | 45 |
| Albite | ........ | ........ | ........ |
| Muscovite | 0.14 | 0.0035 | 40 |
| Albite | ........ | ........ | ........ |
| Muscovite | 0.10 | 0.0030 | 30 |
| Albite | ........ | ........ | ........ |
| Muscovite | 0.10 | 0.0025 | 40 |
| Albite | ........ | ........ | ........ |
| Muscovite | 0.087 | 0.0023 | 40 |
| Albite | ........ | ........ | ........ |
| Muscovite | 0.066 | 0.0020 | 35 |

Data on hydrothermal microcline from pegmatite are scant, but they do indicate strongly that thallium and rubidium are markedly enriched in this mineral relative to the primary type: in fact, if it can be substantiated by further research, this characteristic enrichment of rubidium and thallium in hydrothermal microcline should serve usefully for differentiating between common (primary) pegmatitic microcline and the later hydrothermal type.

Cognizance has, of course, to be taken of possible variation in the general abundance of rubidium and thallium in different petrographic provinces. For example, reference to table 2 shows that specimens of microcline (primary pegmatitic) from the Illovo River area, Natal (nos. 42–45) are unusually poor in thallium and rubidium (mean, 0.0002 per cent $Tl_2O$; 0.01 per cent $Rb_2O$), whereas in the Johannesburg area a much greater concentration of these two elements is found (mean, 0.0009 per cent $Tl_2O$ and 0.1 per cent $Rb_2O$) in microcline of the same type.

*Pollucite.*—Pollucite may be regarded also as a hydrothermal replacement mineral in complex pegmatites, and, in common with all potassium-rich hydrothermal replacement minerals, this cesium mineral shows a characteristic enrichment of thallium and rubidium. The quantities of these two elements found appears to be of the same magnitude as that found in amazonite. In contrast, however, with the potassium minerals, $Tl^+$ and $Rb^+$ substitute for $Cs^+$ in pollucite in place of $K^+$ in potassium minerals.

Enrichment of thallium and rubidium relative to potassium in later hydrothermal replacement minerals accords well with theory: it will be recalled that a geochemical rule enunciated by Goldschmidt (1937) states that when two (or more) ions of like size proxy for a given lattice site in a growing crystal, the smaller (in this instance, $K^+$) ion is accepted more readily because of its greater electrostatic attraction, whereas the larger ions ($Tl^+$ and $Rb^+$) tend to be enriched in the residual mother liquor. This enrichment is extremely marked for cesium (radius $Cs^+ = 1.69$ A) because $Cs^+$ is larger than

$Tl^+$ and $Rb^+$; consequently, cesium may become highly concentrated in residual pegmatitic liquors, thereby causing a precipitation of the cesium mineral, pollucite. This accounts for the fact that, although cesium and rubidium are very similar chemically and although rubidium is much more abundant than cesium, rubidium never forms minerals of its own, whereas cesium does, on occasion, form pollucite.

*Lithium-rich muscovite and biotite from complex pegmatites.*—A marked enrichment of thallium and rubidium is characteristic of lithium-rich muscovite and biotite from complex lithium pegmatites. The enrichment of rubidium had been mentioned by Stevens and Schaller (1942), and Ahrens (1945) has shown graphically in a plot of log percentage $Li_2O$ versus log percentage $Rb_2O$ in muscovite that, although there is a considerable spread in the plotted points, the sympathetic trend in the concentrations of these two elements is quite apparent. Likewise, because of the very close association between thallium and rubidium, lithium and thallium show similar sympathetic trends. In muscovite, thallium and rubidium concentrations vary over considerable ranges.

Data for lithium-rich biotite are very scant, but in the specimen from North Carolina 0.04 per cent $Tl_2O$ and 1.4 per cent $Rb_2O$ were found; this amount of thallium is the highest reported in this investigation. In lithium-rich biotite from the Strickland Quarry, Connecticut, thallium and rubidium were also relatively enriched.

*Common pegmatitic microcline.*—In a high proportion of the specimens analyzed the concentrations of thallium and rubidium fall within limits of 0.01–0.1 per cent $Rb_2O$ and 0.0001–0.001 per cent $Tl_2O$. Because of this wide variation in concentration of thallium and rubidium in these specimens, it is difficult to give accurate mean abundance values: approximate mean values are 0.0004 per cent $Tl_2O$ and 0.04 per cent $Rb_2O$.

The rubidium contents of these microcline specimens appear, in general, to be less than reported elsewhere in the literature; until some interchecking has been done, it will not be possible to ascertain whether the reason lies in the nature of the selection analyzed or whether the discrepancy is due largely to analytical procedure.

In most other minerals, data on thallium and rubidium are very scant and have therefore not been discussed here.

### ABUNDANCE OF THALLIUM IN THE EARTH'S CRUST

Goldschmidt (1937) gives 0.00003 per cent (by weight) as the abundance of thallium in the earth's crust. This value is based on the relatively scant data that were available at the time, and the results of this investigation indicate strongly that this abundance value is too low.

In granite and other rocks relatively rich in potassium, thallium determinations could not be made because the concentration was below the limit of detection of the spectrographic method employed. In arriving at a value for the abundance of thallium in the earth's crust, use may be made, however, of the close association of rubidium and thallium in potassium minerals and of our knowledge of the abundance of rubidium in the earth's crust. According to Goldschmidt (1937), the abundance of rubidium is 0.03 per cent by weight, which value, although possibly subject to some revision because of several more recent analyses, is probably of the correct magnitude. Consequently, using a value of

100 for the ratio $Rb_2O/Tl_2O$, the abundance of thallium in the earth's crust is calculated as 0.0003 per cent, a value which is greater than the older one by a factor of 10.

This value refers to alkali-metal thallium, but it is considered that sulphide thallium present as a trace in some sulphide minerals is insignificant in amount when compared with the thallium content of potassium silicates; consequently, the value of 0.0003 per cent may be regarded as applying to total thallium.

### A COMPARISON OF THE PAIR Rb:Tl WITH OTHER PAIRS OF ELEMENTS

Other pairs of elements are closely associated in minerals: Al-Ga, Rb-K, Ni-Mg, Sr-Ba, Cb-Ta, sundry rare earth elements, and, in particular, Zr-Hf are examples. With the exception of Zr-Hf and possibly some of the rare earths, data on which are very meager, no other pairs of elements are known which are as closely associated as are alkali-metal thallium and rubidium in silicate minerals.

Zirconium and hafnium are very similar chemically and have identical ionic radii; as a result, they have very similar properties, so much so that the discovery of hafnium, which is always present in zirconium minerals, was much delayed. In the common zirconium mineral, zircon, the Zr/Hf ratio has been found to vary only relatively slightly; but in some of the more obscure zirconium minerals the ratio is more erratic. Unfortunately, data are not sufficient to permit a statistical analysis and obtain a normal (probability) curve to compare with the $Rb_2O/Tl_2O$ normal curve. All that can be said now is that the pair Zr:Hf are very closely associated, the closeness of the association being probably similar to that of alkali-metal thallium and rubidium.

The association of zirconium with hafnium differs, however, in some respects from the association of rubidium with thallium; zirconium is a major constituent and forms its own minerals, in which hafnium enters as a minor constituent. Neither rubidium nor alkali-metal thallium forms its own minerals; both elements proxy for potassium in potassium minerals and for cesium in pollucite. Furthermore, like nearly all other pairs of associated elements, zirconium and hafnium are placed close together in the Periodic Table, whereas rubidium is Group 1*a* and thallium is Group 3*b*; hence their very close association in minerals is therefore unique.

## REFERENCES CITED


ADAMSON, O. J. (1942) Minerals of the Varuträsk pegmatite, XXI: The feldspar group: Geol. fören. Stockholm Förh., vol. 64, pp. 19–54.

AHRENS, L. H. (1945) The relationship between thallium and rubidium in minerals of igneous origin: Geol. Soc. South Africa Trans., vol. 48, pp. 207–231.

ERÄMETSÄ, O.; SAHAMA, T. G.; and KANULA, V. (1941) Spektrographische Bestimmungen an Rubidium und Caesium in einigen finnischen Mineralien und Gesteinen: Comm. géol. Finlande Bull. 128, pp. 80–86.

GOLDSCHMIDT, V. M. (1937) The principles of distribution of chemical elements in minerals and rocks: Chem. Soc. London Jour., pp. 655–673.

———; BAUER, H.; and WITTE, H. (1934) Zur Geochemie der Alkali-Metalle: Gesell. Wiss. Göttingen, Math.-phys. Kl., no. 1, pp. 39–55.

STEVENS, R. E., and SCHALLER, W. T. (1942) The rare alkalies in micas: Am. Mineralogist, vol. 27, pp. 525–537.

# REVIEWS

*Fine-grained Alluvial Deposits and Their Effects on Mississippi River Activity.* By HAROLD N. FISK. War Department, Corps of Engineers, Mississippi River Commission, Vicksburg, Mississippi, July 1947. 2 Vols. Pp. 82, figs. 12, pls. 74. $1.25 per volume; $2.50 per set.

The first of these volumes contains the text and figures of the report, the second carries the 74 plates illustrating the text. Presentation of the problem is divided into four parts, which logically develop the author's ideas on the relations of Mississippi River activity to the finer-grained alluvial deposits along the river course.

The Mississippi River and its alluvial valley are discussed in a general way with regard to its physical geology and recent geologic history. Maps are included which show the entrenched valley system during late Wisconsin time and the river course at two later times before the present.

Fine-grained alluvial deposits along the river are described from the point of view of depositional environment, distribution, and physiographic expression. In the second volume a topographic map of the Mississippi River and part of its valley from Commerce Gorge in Missouri to Donaldsonville, Louisiana, is given at a scale of 1:125,000. On this map is shown the distribution and thickness of the fine-grained alluvial deposits as determined by boring and surface mapping. In addition, 359 cross sections are reproduced, showing the thickness and relations of these deposits.

The properties of the fine-grained alluvial deposits are discussed for each type of environment as to grain size, degree of sorting, size distribution within the deposits, and water content. Over fifteen thousand mechanical and hydrometer analyses of samples were made to obtain this information. A more complete list of physical properties is given for the back-swamp deposits, based on detailed testing for the Bayou Sorrell lock site. Detailed compilations of all these data are included in the second volume.

The effect of fine-grained alluvial deposits on Mississippi River activity is described in relation to channel migration, channel cross section, bank caving, stream meandering, and development of reaches.

Effective use is made of block diagrams in illustrating such features as valley development stages and bank caving. Aerial and close-up photographs are used to some extent; but, unfortunately, many of them are not very clear. This is probably due to the fact that they have not been reproduced on glossy paper.

This report is based on very extensive borings and areal mapping done by the Corps of Engineers and contains much authoritative information. It should prove very useful as a source of data on the fine-grained alluvial deposits of the Mississippi River and in supplementing the author's earlier *Geological Investigation of the Alluvial Valley of the Lower Mississippi River.*

A. L. KIDWELL

---

*Geological Explorations in the Island of Celebes under the Leadership of H. A. Brouwer.* Amsterdam: North-Holland Pub. Co., 1947. Pp. 346, figs. 25, pls. 25, sketch maps 2. 27.50 gulden.

This volume is another valuable addition to the geology of the East Indian archipelago to which Dr. Brouwer and his associates have contributed so much. It is divided into three sections: "Geological Explorations in Celebes. Summary of the Results," by H. A. Brouwer; "Igneous and Metamorphic Rocks in Eastern Central Celebes," by W. P. de Roever; "Contribution to the Petrology of the Metamorphic Rocks of Western Celebes," by C. G. Egeler.

Celebes is divided into three main zones: (1) an eastern zone with abundant basic and ultrabasic igneous rocks, schists, Mesozoic and possibly older limestones, and siliceous rocks containing radiolaria; (2) a central zone of epidote mesometamorphic muscovite-rich schists, quartzites, and limestones; and (3) a western zone of granitic to dioritic rocks, gneisses, biotite-rich schists, and Mesozoic and Tertiary volcanic and sedimentary rocks.

The section by De Roever on east-central Celebes is largely descriptive. In a brief petrologic summary on the igneous rocks he concludes that they belong to one differentiation series

comparable with the ophiolite-spilite complex and the Palelo series in Netherlands Timor and to the igneous rocks in the Ligurid overthrust sheet of the Italian Apennines. The metamorphic rocks are classified into the glaucophane-schist and the epidote-amphibolite facies. The glaucophane schist facies is subdivided into two subfacies, one characterized by the stability of lawsonite and the instability of epidote and garnet (lawsonite-glaucophanite subfacies) and the other having the critical associations alkaliamphibole-epidote and alkaliamphibole-garnet with lawsonite unstable. The epidote-amphibolite facies is believed to represent a metamorphism which is older than the radiolarites and the ophiolite and spilitic igneous rocks; the glaucophane schist facies represents a younger metamorphism. Evidence is presented that pumpellyite belongs to a separate metamorphic facies "younger than the radiolarites and igneous rocks, but older than the glaucophanite metamorphism."

The section by Egeler on the rocks of western Celebes contains detailed descriptions of a great variety of metamorphic rocks. He groups these into four facies: greenschist, epidote-amphibolite, amphibolite, and granulite. The amphibolite facies predominates, and the author distinguishes three subfacies, with staurolite, andalusite, and sillimanite-cordierite. Egeler concludes that there were two periods of metamorphism, an older regional upon which was superimposed a younger plutono-metamorphism. The latter is characterized by granitic and granodioritic intrusions with associated intense injection activity and intermingling of igneous and contact rocks. A later low-grade dynamic metamorphism affected the granites.

The authors point out that the glaucophane metamorphism, in eastern Celebes is believed to belong to the same orogenic cycle as does the amphibolitic plutono-metamorphism in the western part. The glaucophane metamorphism and the granodiorite intrusion in the east are late Mesozoic to Tertiary, whereas the epidote-amphibolite metamorphism is older. The regional metamorphism and thermal metamorphism were separated by a long period of sedimentation and volcanic activity. The older group of regional metamorphic facies in the western part is correlated with the epidote-amphibolite metamorphism in the east. The regional metamorphism may have been accompanied by plutonic activity, i.e., some granite and granodiorite may represent this activity.

In his treatment of the structural geology of Celebes, Brouwer states: "From the central part of the Celebes geanticlines branch out to form the four peninsulas, which are separated by deep sea-basins." East of the median zone in Celebes the main strike of minor folds is north-northeast and north-northwest; near the median zone the strike is northeast and east-northeast. The major fractures and faults trend roughly north-south, except in the northeastern part, where they are, in general, east-west. Here the linear arrangement of active and extinct volcanoes shows that they are located on or near lines of fractures. Recent faulting has occurred, but these faults may have been active during long periods, and their age may be considerable.

Major uplift of the peneplane of central Celebes probably "took place in the youngest geological past. Subsidence in fault troughs, which were already in existence in Tertiary time, would then have preceded the general uplift." Brouwer discusses the theories of origin of the Celebes fault troughs and rift valleys.

The book is well printed and bound and has excellent illustrations, in particular the photomicrographs. It is to be regretted that, because of the German occupation, the authors were unable to obtain chemical analyses. Consequently, the parts of the book on petrologic interpretations and comparisons with similar areas are limited in scope. The book, nevertheless, is a worth-while contribution to the geology of Celebes, in particular the metamorphic and structural geology. As Brouwer states, the work is largely "an outline of a field for geological research, which is full of promise."

A. F. HAGNER

---

*Guide to the Geology of Central Colorado*. (*Colorado School of Mines Quart.*, vol. 43, no. 2.) 1948. Pp. 176. $3.00.

This volume contains the work of a number of contributors and is published in conjunction with the Rocky Mountain Association of Geologists. It contains road logs and articles covering the principal geologic features in connection with three field trips arranged for the American Association of Petroleum Geologists. The trips all start from Denver and include: (1) a one-day trip through Boulder, Golden, and Morrison; (2) a two-day trip along the Front Range to Colorado Springs (first day) and to Canyon City and Royal Gorge (second day); (3) a three-day trip along the Front Range, Mosquito-Gore

Range, Glenwood Canyon to Glenwood Springs (first day), to Rifle Gilman, Red Cliff, and Leadville (second day), and along the Arkansas Valley, across the Mosquito Range to South Park and Denver (third day).

The articles cover the general geology of central Colorado, the Kokomo, Gilman, Leadville, and Climax mining areas, and the oil-shale deposits near Rifle. Most of them are summaries or revisions of published papers. The United States Bureau of Mines oil-shale demonstration plant at Rifle is described. The volume contains thirty-eight geologic and columnar sections, maps, diagrams, and photographs. A map gives the routes of the trips, the major geologic structure, and oil and gas information. The bulk of the volume is devoted to road logs and to the stratigraphic and structural geology of central Colorado; it serves as a useful guide to the region.

A. F. HAGNER

---

*Eruptive Rocks: Their Genesis, Composition, Classification, and Their Relation to Ore Deposits, with a Chapter on Meteorites*. By S. J. SHAND. 3d ed. New York: John Wiley & Sons; London: Thomas Murby & Co., 1947. Pp. xvi+488; figs. 51; pls. 4. $7.50.

The first edition of Professor Shand's book (1927) marked a new approach to many petrological problems and soon established itself as a particularly valuable guide for the novice as well as the expert.

In the third edition the content has been rearranged, in that the chapters of modal analyses of rocks, which are intended to illustrate Shand's well-known system of classification, have been moved to the end of the book, where they form Part II. This is an improvement. The book is now exceptionally well written, interesting, and logical in its setup.

Three new chapters have been introduced to bridge the gap between eruptive rocks and certain types of ore deposits. Much valuable information has been collected. The subject matter, including such highly controversial problems as the genesis of pegmatites and late magmatic and postmagmatic reactions, is logically and clearly presented. There are no wild speculations, each statement is carefully considered and supported by thoroughly checked facts.

Professor Shand's superior insight and intimate knowledge of the eruptive rocks makes this new edition a source of information of great value to our science.

However, when reading this book, a rather fundamental question in the teaching of petrology insinuated itself into my thoughts. A couple of examples will illustrate the problem.

Shand's discussion of the order of crystallization is purely descriptive; no effort is made to present and use the elementary laws of physical chemistry. In my opinion it is much easier for the student to understand and remember the facts if they are logically related to fundamental natural laws. Shand completely avoids any use of mathematical symbols; thus his descriptions become so qualitative as to be inaccurate. It is *not* true that "leucite is compatible with free silica above 1170° C" (pp. 99 and 122); and the "simplified" diagram of $MgO$–$SiO_2$ (p. 223) is radically wrong. The phenomenon of incongruent melting, so fundamental in petrogenesis, cannot be understood at all from such oversimplified statements.

Likewise, the section on geologic thermometry (chapter on temperature and pressure in the magma) is purely descriptive and conspicuously lacking in quantitative background. Much space is devoted to telling what this and that geologist thought about the influence of pressure on the inversion-point quartz-tridymite; but the reader is never taught that there is a very important thermodynamic relation concerning the equilibrium between phases called "Clapeyron's equation."

Are petrology students really so deficient in general scientific background as to be unable to appreciate the great simplification introduced by a thermodynamic equation? I do not think so. And if some students are, they should be required to make up for such deficiencies.

Professor Shand has a fascinating way of teaching the petrology of eruptive rocks. Outdated notions are left behind like dead wood, and a treatment has been introduced which is, to a large extent, based on the chemistry of silicate systems. Yet nobody can adequately teach the chemistry of such systems without making use of the fundamental laws of thermodynamics.

T. F. W. B.

---

*Biography of the Earth*. By GEORGE GAMOW. New York: New American Library, 1948. Pp. xiii+194. $0.35.

This is a pocket edition of the popular book, first published in 1941, with the addition of a section to the chapter on the origin of the earth.

The volume is directed at the nonscientific reader, and the author does a good job of putting into everyday language the astronomical and geological facts of the story. Geologists will find minor points on which they will disagree with the author (as well as among themselves) and will feel that he tends to put too much reliance on deductions drawn from facts which may themselves be poorly established. On the whole, however, the book is accurate as well as entertaining. The Preface indicates that misstatements and misprints in the original edition have been corrected, but a few still remain. An example is the statement that the term "Cambrian" comes from the eastern England county, instead of Cambria, Wales. Erosion is stated to have removed a 4-inch layer from the surface of the United States since Columbus landed in the New World, although the rate of denudation is given (correctly) as about 0.02 mm. per year —or $\frac{3}{8}$ inch since 1492.

American readers would prefer American rather than English spellings where these differ, and the reviewer feels that well-chosen references to guide the reader further into many of the fields touched upon would be valuable. The author succeeds in rousing an interest in every phase of his subject, which ought to be encouraged by suggested readings.

M. S. C.

---

*Gem Testing.* By B. W. ANDERSON. New York: Emerson Books, 1948. Pp. 256; figs. 53. $5.00.

The purpose of this work is to serve as "a simple yet detailed manual for jewelers and others who are interested in the identification of gem stones." The earlier chapters cover refractometry and immersion index methods, the dichroscope, specific gravity, the microscope, and the spectroscope. There is a color plate showing absorption spectra. The remainder of the book deals with the identification of the various gem stones, synthetic and natural. An appendix contains a glossary of terms, lists of properties, etc. There is an index. There are a number of good photomicrographs, and the technical job of bookmaking is well done. The book would seem to fulfil its objective in praiseworthy fashion; it is recommended for the serious nonspecialist.

D. J. F.

---

*Gems and Gem Materials.* By E. H. KRAUS and C. B. SLAWSON. 5th ed. New York: McGraw-Hill Book Co., Inc., 1947. Pp. ix+332; figs. 403. $4.00.

The arrangement of the contents of this highly successful elementary book has not changed since it was first published in 1925. So many minor alterations have been made that this edition has been reset. There is now a brief bibliography, as well as a new eight-page section on "Crystal Structure and X-Ray Methods." The fine color plates taken from Eppler which were present in the third edition are missing.

D. J. F.

## ERRATA

In "Deformation of Quartz Conglomerates in Central Norway," by Christoffer Oftedahl, in the September 1948 *Journal*.

p. 481, eq. (4) read $\left.\begin{matrix}a\\c\end{matrix}\right\}$ instead of $\left.\frac{a}{c}\right\}$

p. 486, eq. (14) read $\left.\begin{matrix}a\\c\end{matrix}\right\}$ instead of $\left.\frac{a}{c}\right\}$

col. 2, line 17, read *b* instead of b

p. 487, col. 1, line 8, read *b* instead of b

line 16, read *a* and *c* instead of a and c

col 2, line 3, read *b* instead of b

line 10, read *b* instead of b

line 13, read *b* instead of b

# INDEX TO VOLUME 56